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PAPERS IN PRESS FOR FORTHCOMING ISSUES

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SECRETARY'S NOTE

Separates of the *Geochemical table of the elements for 1959*, accompanying the paper by Jack Green in the September issue, are available at $1.00 per copy (unfolded, in tubes). Please send orders with remittance to The Geological Society of America, 419 West 117 Street, New York 27, N. Y.

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
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THICKNESS AND CONSOLIDATION OF DEEP-SEA SEDIMENTS

BY EDWIN L. HAMILTON

ABSTRACT

Geochemical and other calculations on the total volume of sediment that should be present in deep-sea basins, based on materials released in weathering, indicate that the average thickness of sediment solids on the ocean floor should be 1–3 km. Seismic surveys in all oceans, on the contrary, report a sediment thickness on the order of 0.1–0.5 km but also reveal that, in many areas, the surface sediments are underlain by other layers whose elastic-wave-transmission velocities are intermediate between those of sediment and the "basaltic basement" rock of the oceanic crust.

Examination of the above problem in the light of soil-mechanics studies, and recent studies of the velocity of elastic compressional waves in sediments and rocks, indicates the possibility that, in many areas, the lower layers are composed of rocks formed by normal consolidation and lithification of the present types of deep-sea sediments.

Gravitational consolidation of red clay should reduce sediment pore space to values between 35 and 45 per cent at depths in the sediment between 150 and 700 m; between these porosities, pressure-chemical effects in the dewatered sediment should result in lithification to shale and marked increase in seismic velocity.

Gravitational consolidation of calcareous ooze should result in porosity reduction at a slower rate than in clay, but lithification of this material with age and pressure can result in marked seismic-velocity increases at or near the sediment surface; the discovery of deep-sea limestone in the form of lithified *Globigerina* oozes ranging in porosities from 5 to 68 per cent and in compressional elastic-wave velocities from 1.8 to 5.5 km/sec indicates the possible velocity variations. Various combinations of clay and calcareous material are possible and probable in many areas; high velocities under a thin cover of present-day clay deposition thus possibly represent an ancient (Cretaceous to Tertiary) layer of lithified calcareous ooze.

The present sediment thickness measured in deep-sea basins is probably the total thickness of these upper layers. If so, the thickness of all layers above the "basalt" should enter the calculation of total sediment deposited on the sea floor. A measurement of sediment thickness at any time horizon of the past or future should show about the same "anomalously thin" value for the topmost part of the slowly consolidating and lithifying sediment.

Normal, expectable deposition of sediment in many deep-sea areas, and subsequent consolidation and lithification, may explain most of the seismic discoveries in these areas and fall into line with the great geochemical balance and with the theories of soil-mechanics engineering.

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Introduction

Refraction and reflection seismic exploration began in the deep sea in the 1930's, but it was not until after World War II that investigators began to get a general idea of sediment thicknesses and crustal structure in the ocean basins.

Before the beginning of seismic work in deep-sea areas many estimates were made of the total thickness of sediments on the sea floor above the topmost crustal layer, which was usually presumed to be basalt. Kuenen summed up these estimates in 1950 (p. 386–398). Approaching the question from five different standpoints Kuenen deduced that the deep-sea floor should have an average thickness of about 2½ km of solids or about 3 km of sediments (including pore space). Revelle (1954, p. 327) noted that the geochemical evidence required 1–3 km of sediments on the sea floors.

It was a surprise to marine geologists, therefore, when the first results of seismic-refraction work in deep-sea areas indicated that only 200–500 m of unconsolidated sediments overlay material thought to be basalt of the earth's outer crust, as indicated by the velocity of the compressional wave and by petrological considerations. This "anomalous thinness" of sediments immediately became one of the important problems of marine geology.

Several reasons were advanced to explain this thin veneer of sediment over basalt: (1) the lower part of the sedimentary column had disappeared as a result of some catastrophic occurrence, (2) the lower part of the column was undetectable by seismic means, (3) vast floods of lava or other volcanic rocks were spread over the sea floors, so that high velocities were obtained just under the unconsolidated sediments or (4) the rate of deposition had been extremely low.

After these early seismic estimates of sediment thickness reached the literature, improved techniques of seismic work at sea and interpre-

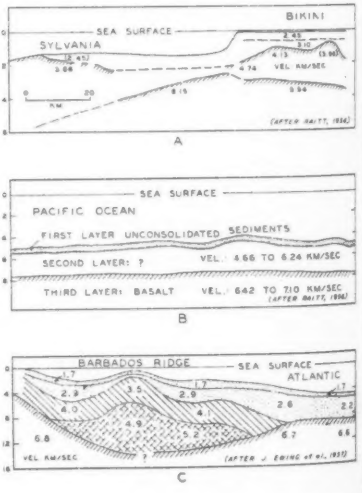

FIGURE 1.—LAYERED STRUCTURE OF THE SEA FLOOR IN SELECTED AREAS

A: North-south section through Bikini and Sylvania (after Raitt, 1954); B: diagrammatic structure in the northeast Pacific Basin (after Raitt, 1956); C: east-west generalized structure across the Barbados Ridge (after J. Ewing *et al.*, 1957)

tation of the records began to show that a "second layer" or multiple layers occurred between the unconsolidated sediment and the "basalt" layer in many areas (Officer, 1955; Raitt, 1956); Hersey *et al.* (1952) showed an example of thicknesses and layering in the Western Atlantic under differing assumptions (one of which assumes no second layer). These multiple layers were noted in most shallow-water work almost from the beginning, but only in the past few years have they been noted consistently in the deep-sea areas. Raitt was one of the pioneers in the deep-sea determinations, and his papers (1954; 1956) show second layers throughout the Pacific Basin as well as multiple layering on the tops of seamounts. Figure 1A illustrates the layering that Raitt found (1954) on the top of Sylvania seamount in the Marshall Islands; Figure 1B is a generalized composite of Raitt's results from the northeast Pacific.

A summary of the layers above the basalt for several areas in the Pacific follows (values averaged from Raitt, 1956):

Area	First layer		Second layer		Total thickness (layers 1 and 2) (km)
	Thickness (km)	Velocity (km/sec)	Thickness (km)	Velocity (km/sec)	
NE. Pacific red clay†	0.29	2.15*	0.82	5.91	1.11
E. Pacific calcareous**	0.33	2.15*	1.06	5.29	1.39
Apron, Mid-Pacific Mountains††	0.39	2.15*	2.33	4.86	2.72

* Assumed

† Mid-Pacific stations 1, 2, 3, 4, 6, 7

** Capricorn stations 18, 19, 20, 21, 22, 23, 24, and 25; Mid-Pacific station 5

†† Mid-Pacific stations 10, 11, 12

Fisher and Shor (1956), after seismic work on the shelf off the west coast of Guatemala, noted five layers with velocities ranging from 1.71 km/sec at the surface to 5.96 km/sec just above the main crustal layer; they thought that the first three layers (1.71 to 3.25 km/sec) probably represent sediments of varying stages of compaction. Seismic work on the continental shelf southwest of England by Hill and Laughton (1954) indicated layered structure which was thought to be consolidating sediments and sedimentary rock.

Multiple layering has been noted in the Gulf of Mexico-Caribbean area and parts of the deep Atlantic by exploratory parties from the Lamont Geological Observatory and the Woods Hole Oceanographic Institution. Excellent examples of this are noted by Hersey *et al.* (1952), M. Ewing *et al.* (1955), Officer (1955), Officer *et al.* (1957), J. Ewing *et al.* (1957), and Ewing and Ewing (1959). An east-west structure profile from J. Ewing *et al.* (1957) is illustrated in Figure 1C.

Early analyses of the seismic results from the cruise of HMS CHALLENGER (1950–1953) did not indicate layering in the deep-sea areas, but re-evaluation of these results demonstrated that layering is typically present, even in the deep-sea areas (Gaskell, Hill, and Swallow, 1958). An example of the new interpretation is Challenger Station 8 in the northeast Pacific (Lat. 35° 01′ N., Long. 143° 14′ W.), where, in 2900 fathoms of water, a first layer 0.1 km thick overlies a second layer 0.6 km thick in which the seismic velocity is 4.5 km/sec.

Hill (1957), in a masterly summary of recent geophysical exploration in the oceans, concluded that evidence against the existence of layer 2 is not good, and that in numerous places evidence for its existence is good. He was strongly in favor of a layer 2 between the sediment and basalt. Hill thought that those stations where layer 2 has not been recorded should be eliminated from averaged thicknesses of layer 1, the unconsolidated sediment layer, because, if layer 2 does exist, then the calculated thicknesses of layer 1 are excessive.

The material of the second layer has been thought to be granite, lava flows, pyroclastic rocks, metamorphic rocks, consolidated sediments, and mixtures of these rocks and others of appropriate velocity; most workers favor volcanic rocks. Around continental margins and near islands, or former islands, almost all these rock types or various combinations of them are possible. In the deep-sea areas, however, the situation should be simpler. The writer (1957) favored lithified sediment as the second-layer material in the deep sea, as do Gaskell *et al.* (1958). Hill (1957) thought that the second layer in deep-water areas was either volcanic rocks or consolidated sediment, or a combination of the two.

This paper discusses the consolidation and lithification of deep-sea sediments, the velocities of compressional elastic waves expectable as a consequence of such changes, and the probability that the second layers, in many areas, are lithified deep-sea sediments. A major conclusion of this paper is that normal, expectable

deposition of sediments in many deep-sea areas, and subsequent consolidation and lithification, explains most of the present seismic discoveries in these areas and falls into line with the great geochemical balance and with soil-mechanics engineering theory.

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Consolidation and Lithification of Deep-Sea Sediments

Consolidation Theory

The consolidation of soils and sediment has received more than casual interest during the past 4 decades. In 1923 Terzaghi concluded a study of consolidation which is the cornerstone of soil mechanics (*see* Terzaghi, 1943); subsequently this subject has been intensively studied by foundation engineers because the selection of proper foundations is directly dependent on proper evaluation of the amount of settlement expectable under a given design load (Terzaghi and Peck, 1948; Taylor, 1948).

A saturated sediment is a porous structure of mineral and/or rock grains. When a natural or man-made load is applied to this structure the volume decreases because of the escape of water from the pores. The speed of this volume reduction is primarily dependent on the amount and rate of load application and the permeability of the sediment.

The pressure changes in a water-filled sediment are analogous to those in a water-filled cylinder fitted with a spring and a piston through which a stopcock allows water to escape (Taylor, 1948, p. 223). When a load is applied to the piston and the stopcock is closed, all the load is borne by the water (the water being much less compressible than the spring); when the stopcock is opened, the load is gradually transferred to the spring from the water as the volume of water decreases; finally the load is entirely borne by the spring. The pressures in a sediment act in much the same way except that pressures are the same throughout the height of the cylinder; in a sediment over an impervious base the consolidation begins at the top and progresses downward (Taylor, 1948). A load suddenly applied is at first carried by the pore water (as excess pore-water pressure) and gradually transferred to the sediment structure; the consolidation results from the decrease of pore space through grain repacking and the expulsion of water. The primary consolidation is completed when the excess pore-water pressure is zero and pore-water pressure is solely hydrostatic. A secondary consolidation may continue for some time, probably because of the plastic deformation of grains and of the adsorbed water around the grains (Taylor, 1948; Peck, Hanson, and Thornburn, 1953).

When the rate of load application is very slow Taylor (1948) thought it possible that minute loads cause almost no excess pore-water pressure, and consolidation, therefore, is almost all secondary. It appears that the extremely slow deposition on the deep-sea floor resulting in minute load applications causes little excess pore-water pressure.

The term "consolidation" is used in soil mechanics to refer to a process and not to the solidity of material. A material can be completely consolidated (under a certain load, or overburden) and still be relatively soft. Geologists should not use "consolidated sediment" when they mean "lithified sediment" or "rocks." In engineering parlance "compaction" is reserved for artificial reduction of pore space as with a sheepsfoot roller.

The consolidation behavior of a clay is studied by running a consolidation test in the laboratory on relatively undisturbed samples and on remolded samples of the same sediment. The sediment is placed in a ring with porous discs at top and bottom. A load is applied to the top disc, and the water is permitted to flow through the discs until volume reduction of the sediment has practically ceased. The amount of reduction is measured on a dial gage. When the rate of volume reduction becomes very small a new load increment is added.

The results of a consolidation test are plotted as the void ratio, e (volume of voids divided by the volume of solids), versus the corresponding total pressure, p. When plotted on semilogarithmic paper the curve is called, in soil mechanics, an "*e-log p*" curve and is used in the analysis of settlement that might be expected under given design loads.

After overburden and other load pressures are computed for an actual condition then one can enter the *e-log p* plot along the pressure line and pick out the corresponding void ratio

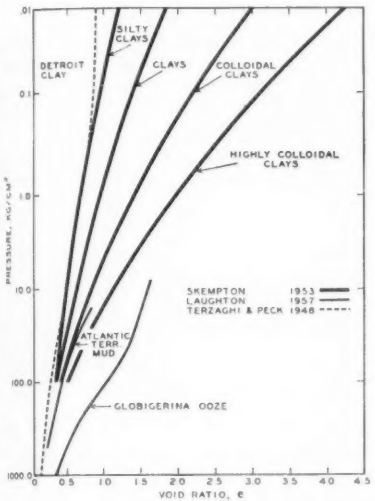

FIGURE 2.—CONSOLIDATION (*e-log p*) CURVES FOR CLAY AND *Globigerina* OOZE

for the particular pressure. Examination of a number of *e-log p* plots for clay reveals a very significant and important fact: no matter how porous the original material, by the time it is under about 100 kg/cm² of pressure the void ratio has usually decreased to about 0.3–0.5 (porosity approximately 29 per cent). Figure 2 illustrates this fact and shows the funnelling effect derived when a number of plots are made for different clays; this convergence is even more marked at pressures greater than about 100 kg/cm². The curves from Skempton (1953), shown in Figure 2, are derived from a large number of field and laboratory determinations which are in accord with a large number of plots from other sources collected by the writer but not shown on the figure to preserve clarity. Laughton (1957) has published results of consolidation-velocity studies on deep-sea materials; in Figure 2 his *e-log p* curves have been added to the curves of Skempton and Terzaghi and Peck (1948).

How close do the laboratory determinations come to the actual consolidation in the field (settlement under buildings or under gravitational consolidation due to the weight of the sediment)? This matter has been extensively studied, the laboratory results allow close prediction of actual consolidations (Skempton, 1953; Schmertmann, 1955). Usually the laboratory-derived void ratios are slightly less than those in the field at a given pressure; this is thought to be due to loss of strength because of disturbance during sampling.

The intergranular pressure (due to the weight of sediment grains) used in computations of consolidation is independent of the depth of the water as long as free movement of the water can take place so that the excess pore-water pressure can become zero. The question of hydrostatic uplift (buoyancy of mineral particles due to Archimedes principle) is important in computing pressures within the sediment. In the field of soil mechanics 100 per cent uplift is allowed in foundation studies. In the present study all pressures are computed with full hydrostatic uplift in effect; this is the minimum case for explaining high seismic velocities, because velocities are likely to be higher with higher pressures. The question of hydrostatic uplift is considered in more detail in Appendix A.

Gravitational Consolidation of Deep-Sea Sediments

Computations.—Excellent summaries of the subject of deep-sea sedimentation have been published by Revelle *et al.* (1955) for the Pacific, and by Ericson *et al.* (1955) for the Atlantic. The two dominant types of sediments are red clay in the deeper areas and calcareous ooze in the shallower areas (less than about 2400 fathoms) of low and mid latitudes. Arrhenius has shown (1952) that high productivity in the upper waters may lead to more extensive deposits of calcareous material than would normally be expected. Calcareous ooze is formed largely by the empty shells, or tests, of the protozoan Order Foraminifera and by coccolith plates and is thus a calcareous sand or sandy silt of a specialized type.

In constructing Table 1, which shows the probable consolidation of a red-clay shale, the writer used the *e-log p* curve for colloidal clay (Fig. 2) from 0.1 to 19.5 kg/cm² (where the porosity is 41 per cent and sediment depth is 300 m) and then the curve for Skeel's shale with hydrostatic uplift, at greater pressures; deep-sea clays and silty clays will vary only slightly from these determinations in the upper portion of the geologic column. In computing Table 2 for *Globigerina* ooze the *e-log p* data from Laughton (1957) were used to illustrate

the consolidation of this typical deep-sea sediment. After tables have been computed it is then possible to construct various curves to illustrate the interrelationships of the various properties (Figs. 3, 4).

Standard soil-mechanics computations allow the prediction of settlement in a sediment section, provided the void ratios before and after a consolidation test are known (*e.g.*, Tschebotarioff, 1951, p. 105). Using the same formulations it is possible to expand any depth increment of a sediment column to show how much sediment was originally present to consolidate to the stated increment. The arbitrary increments of Table 1 are thus expanded and shown in Table 4 (Appendix B). The amounts of clay necessary to consolidate to present thicknesses are surprisingly large: 1000 m of original deposition will consolidate to about 500 m; it took more than 5000 m of clay to form a present consolidated section 2000 m thick.

TABLE 1.—CONSOLIDATION OF CLAY AND SHALE

Assumption: 300 m of clay overlies shale with 100 per cent hydrostatic uplift effective. Properties: density of solids, 2.31 gr/cc; water density, 1.05 gr/cc; initial porosity, 72 per cent.

Depth in sediment (m)	Interval pressure (Kg/cm²)	Cumulative pressure (Kg/cm²)	Void ratio, e	Density saturated (gr/cc)	Submerged weight (gr/cc)	Porosity n (per cent)
0	0	0	2.58	1.40	0.35	72
10	0.35	0.35	1.73	1.52	0.47	63
20	0.48	0.83	1.48	1.55	0.50	60
30	0.51	1.34	1.35	1.58	0.53	58
40	0.54	1.88	1.25	1.60	0.55	56
50	0.56	2.44	1.18	1.63	0.58	54
60	0.59	3.03	1.13	1.64	0.59	53
70	0.60	3.63	1.08	1.66	0.61	52
80	0.62	4.25	1.05	1.67	0.62	51
90	0.63	4.88	1.02	1.68	0.63	50
100	0.63	5.51	0.98	1.69	0.64	49
150	3.3	8.8	0.87	1.72	0.67	47
200	3.5	12.3	0.80	1.76	0.71	44
250	3.6	15.9	0.75	1.77	0.72	43
300	3.7	19.6	0.70	1.80	0.75	41
400	7.7	27.2	0.58	1.85	0.80	37
500	8.1	35.3	0.54	1.87	0.82	35
600	8.3	43.6	0.50	1.90	0.85	33
700	8.6	52.2	0.47	1.91	0.86	32
800	8.7	60.9	0.43	1.93	0.88	30
900	8.9	69.8	0.40	1.95	0.90	29
1000	9.1	78.9	0.37	1.97	0.92	27
1500	48.4	127.3	0.25	2.06	1.01	20
2000	52.8	180.1	0.15	2.15	1.10	13

TABLE 2.—CONSOLIDATION OF A *Globigerina* OOZE

Assumptions: A progressive porosity decrease under pressure with no lithification which is probably a rare situation (see text). Properties: Density of solids: 2.69 gr/cc; water density: 1.05 gr/cc; initial porosity: 66 per cent.

This table is derived from Laughton (1957) who ran the test on a *Globigerina* ooze with 54 per cent $CaCO_3$.

Depth in sediment (m)	Interval pressure (Kg/cm²)	Cumulative pressure (Kg/cm²)	Void ratio, e	Density saturated (gr/cc)	Submerged weight (gr/cc)	Porosity, n (per cent)
0	0	0	1.90	1.62	0.57	66
10	0.57	0.57	1.88	1.62	0.57	65
20	0.57	1.14	1.85	1.63	0.58	65
30	0.58	1.72	1.82	1.63	0.58	65
40	0.58	2.30	1.80	1.64	0.59	64
50	0.59	2.89	1.78	1.64	0.59	64
100	3.00	5.89	1.68	1.66	0.61	63
200	6.2	12.1	1.56	1.69	0.64	61
300	6.5	18.6	1.49	1.71	0.66	60
400	6.7	25.3	1.44	1.72	0.67	59
500	6.8	32.1	1.40	1.73	0.68	58
600	6.9	39.0	1.36	1.74	0.69	58
700	7.0	46.0	1.32	1.76	0.71	57
800	7.2	53.2	1.27	1.77	0.72	56
900	7.3	60.5	1.23	1.79	0.74	55
1000	7.4	67.9	1.18	1.80	0.75	54
1500	39.0	106.9	1.01	1.87	0.82	50
2000	42.5	149.4	0.88	1.92	0.87	47

The above technique is extremely important in determining, approximately, how much total sediment is represented in any given column and is an entry to studies of the amounts of solids in deep-sea sediments and ages of ocean basins. These matters will be the subject of a later paper.

The values used for the physical properties of the sediments of Tables 1 and 2 are shown below each table; these were the starting values for the sediment at the surface of the present-day sea floor. The methods of computation used in Tables 1, 2, and 4 are discussed in Appendix B.

Discussion.—The difference between the porosity of a calcareous ooze and a clay at the same pressure and depth is distinctive: the clay

porosity is less than that of calcareous ooze (Fig. 2).

The different consolidation behavior of sand and clay has long been known and studied in the field of soil mechanics. Sands, after a quick mechanical rearrangement of particles, will show relatively slow decreases of porosity under increasing pressures. The adsorbed water layer, if any, around sand grains is unimportant in this consolidation history. Clay shows a relatively fast decrease in porosity with increasing pressure and with expulsion of both the pore water and the adsorbed water around the grains. Figure 5 illustrates these differences at pressures greater than 50 kg/cm². Mixtures of sand and clay will have consolidation curves intermediate between clay and sand but always on the high porosity side of clay at the same pressure.

Globigerina ooze is largely composed of shells, or tests, of planktonic Foraminifera; mechanical analyses of typical calcareous oozes (Revelle, 1944, p. 144–147) show that the size distribution of particles is such that most calcareous oozes can be considered as very specialized types of sandy silt or sand-silt-clay. Therefore, the behavior of *Globigerina* ooze, or calcareous ooze, under gravitational pressure should follow the expectable consolidation curve for sandy silt or silty sand. In addition, calcareous ooze is a specialized sandy silt in that many of the grains are hollow shells of calcium carbonate; these delicate shells after crowding together under compaction will crush much more easily than typical shallow-water sandy silt grains. Thus calcareous ooze should show a faster decrease in porosity than typical quartz-particle sediments; Figure 5 illustrates this for the ooze that Laughton studied, and it should be true of other similar oozes of high calcareous shell content.

The difference between the porosity of deep-water sediments (depths greater than 1500 fathoms) and shallow-water sediments (depths less than 100 fathoms) was discussed by Nafe and Drake (1957) who show that, in the Atlantic, deep-water sediments have a much

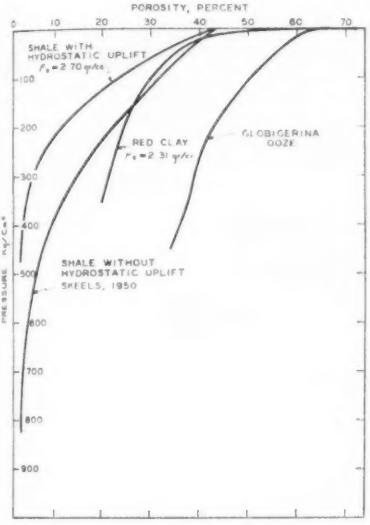

FIGURE 3.—POROSITY VS PRESSURE RELATIONSHIPS FOR RED CLAY, *Globigerina* OOZE, AND SHALE WITH AND WITHOUT HYDROSTATIC UPLIFT OF SOLIDS

Porosity and pressure data for shale from Skeels (*in* LeRoy, 1950); *see* Tables 1, 2 for data on sediments.

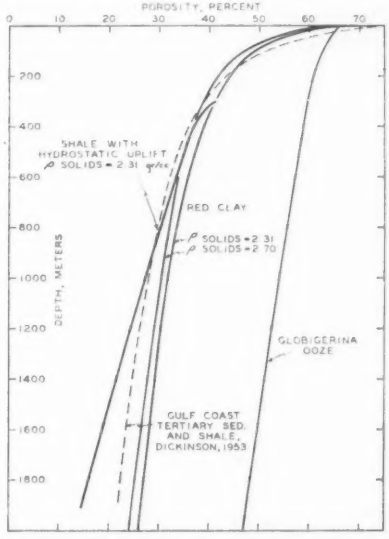

FIGURE 4.—POROSITY VS DEPTH BELOW THE SEA FLOOR FOR RED CLAY, *Globigerina* OOZE, AND SHALE

Data on clay and *Globigerina* ooze from Tables 1, 2; data for shale with hydrostatic uplift derived from Skeels' data (*in* LeRoy, 1950) assuming 300 m of unlithified clay overburden, hydrostatic uplift, and an average density for Pacific red clays; data on Gulf Coast Tertiary sediments and shale from Dickinson (1953) with no assumptions.

higher porosity than shallow-water sediments at the same depth of burial. This analysis, however, is complicated by the fact that much of the Atlantic continental terrace beneath the shallow-water sediments is composed of various kinds of hard rocks, some of which may have been eroded. In the uppermost parts of these two sections, the fine silts and clays of the continental shelf should decrease in porosity with depth at a faster rate than the deep-sea sediments which in many areas are likely to contain foraminiferal tests and sands and coarse silts laid down by turbidity currents; thus, at least in the topmost sections, the differences in porosity are likely to be due to the differing consolidation characteristics of the sediments.

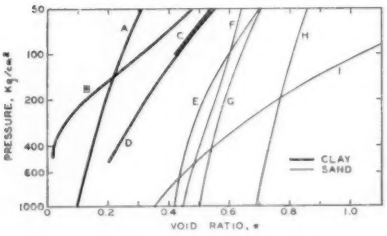

FIGURE 5.—CONSOLIDATION (*e-log p*) CURVES FOR CLAY AND SAND

A: soft Detroit clay; B: shale with hydrostatic uplift; porosity data from Skeels (*in* LeRoy, 1950); C: colloidal clay (Skempton, 1953); D: terrigenous mud (Laughton, 1957); E: 80 per cent sand and 20 per cent mica; F: dense sand; G: 90 per cent sand and 10 per cent mica; H: loose sand (A and E-H from Terzaghi and Peck, 1948); I: *Globigerina* ooze (Laughton, 1957)

Figure 4 indicates only slight differences in consolidation characteristics between a clay with a density of mineral particles of 2.70 gr/cc (the highest expectable) and one with a density of 2.31 gr/cc (the average of 18 determinations on Pacific red clay); the heavier particles cause only a slightly greater decrease in porosity to any given depth.

Lithification of Clay and Calcareous Oozes

Clay.—From 1925 to 1935 many people studied the process by which a silt or clay is consolidated and lithified to form a shale. Hedberg (1936) published the classic study of the subject to date. He used a combination of soil-mechanics knowledge and a field study of undistorted shales in Venezuela to investigate the relations between pressure, depth, porosity, density, and lithification. Hedberg mentioned at the start of his study that there would never be an exact quantitative solution to the relations between pressure, volume, and lithification in clays and shales because too many variables were involved, but that generalizations were possible for the argillaceous sediments. Figure 3 of this paper is similar to Figure 5 (p. 262) of Hedberg's paper with the red-clay curve replacing Hedberg's curve from Terzaghi's early work, and with Skeel's data (*in* LeRoy, 1950, p. 340) replacing his pressure porosity for shales. Hedberg's ideas are illustrated in Figure 6 (from Hedberg, 1936, p. 281, Fig. 7); briefly summarized his concepts are as follows:

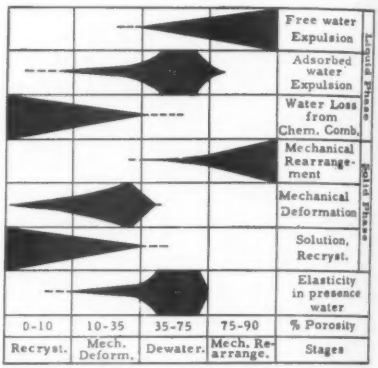

FIGURE 6.—PROCESSES IN THE CONSOLIDATION AND LITHIFICATION OF CLAY
(From Hedberg, 1936, Fig. 7)

(1) Mechanical rearrangement stage (90–75 per cent porosity): As a highly porous clay, with its particles surrounded by adsorbed water, is placed under gravitational pressure the pore water, or free water, is first squeezed out as a consequence of the mechanical rearrangement of particles; some adsorbed water is lost.

(2) Dewatering stage (75–35 per cent porosity): The expulsion of free water continues; adsorbed water moves from the points of maximum stress, at the intergrain contacts, into the larger pores. With increasing pressure the approximate lower limit of the stage is reached (at about 35 per cent porosity) when the solid grains begin to come into contact with each other over some part of the surface area; only a small amount of adsorbed water remains, and almost all the free water is gone. During this

stage, if the pressure is removed, Hedberg believed that although there might be some elasticity of grains, the dominant factor causing the "elastic" expansion of the sediment is the reattainment by each particle of its film of adsorbed water. Modern studies (*e.g.*, Lambe, 1953, p. 32) affirm that this concept is essentially correct.

(3) Mechanical deformation stage (35–10 per cent porosity): This stage is reached when the porosity is reduced below about 35 per cent. Individual particles are in contact or near contact, and further consolidation can occur only with the alteration of the shape or structure of the grains. Hedberg thought that this is a mechanical process with bending, crushing, and granulation of the clay particles. More adsorbed water is forced out, and chemical readjustment begins; there is a tendency for pressure-point solution and redeposition in adjacent regions under lower stress; the colloidal matter tends to crystallize with the formation of new minerals.

(4) Recrystallization stage (10 per cent to 0 porosity): Chemical readjustment becomes the dominant factor; total volume is reduced by the formation of denser minerals, and pore space is filled by these new minerals and by cementation. Hedberg thought that this phase starts at about 10 per cent porosity; with further pressure and higher temperature at depth, the larger crystals grow at the expense of the smaller, and if conditions are suitable a gradual transition may occur from a shale to a slate and then to a phyllite.

Comparison of the above stages with some of the latest ideas on consolidation shows that little has been added beyond Hedberg's summation of knowledge in the field.

An interesting and significant question in the above sequence of consolidation is: at what point does the clay become shale? This transition, according to the above hypothesis, apparently comes at about 35 per cent porosity where because of chemical changes cementation between the grains imparts a rigidity to the structure which previously was not present. Skeels (*in* LeRoy, 1950) has noted shales with porosities up to 43 per cent; Birch *et al.* (1942) record a maximum shale porosity of 45 per cent.

The appearance of the porosity vs depth curve (Fig. 4) indicates that the change from clay to shale might come in a minimum depth range of 150–250 m; the porosities at these depths are 45 and 40 per cent. This evidence is consistent with the known highest porosities of shale noted above.

Obviously the change from a clay to a shale will occur at different depths and porosities, depending on pressure and chemical effects, but a reduction to porosities between 30 and 43 per cent is indicated by hypothesis and field data to be the usual range wherein clay becomes shale. These porosities should be reached at depths between 150 and 700 m (Fig. 4), depending on chemical factors and pressure.

Calcareous ooze.—Deposition of calcareous ooze in deep-sea areas probably began during the Cretaceous period with the evolution of the planktonic Foraminifera, although previous deposits of coccolith plates are not excluded (Bramlette, 1958). Calcareous ooze has been accumulating for about 100 million years; today it constitutes about 36 per cent of the areal extent of sea-floor sediments. The consolidation history of this material is thus of some importance in understanding deep-sea sedimentation.

Consolidation data on calcareous ooze are rare. R. D. Terzaghi (1940) ran tests on some limy muds from the Bahamas, and Emery *et al.* (1954) briefly studied consolidation of *Halimeda* deposits. Laughton (1957), however, ran a full, modern, consolidation test on a deep-sea *Globigerina* ooze (Fig. 2). As previously noted, the curves for clay and for *Globigerina* ooze differ significantly: the void ratios for the calcareous material are much greater than for the clays at the same pressures.

A bed or deposit of calcareous material under pressure would be expected to consolidate along the lines shown in Table 2. During this consolidation, after the pressure had crushed many of the foraminiferal shells, or tests, pressure-point solution and reprecipitation would result in both cementation and recrystallization during mechanical and plastic-grain deformation. This cementation and recrystallization would continue after consolidation and result in further reduction of porosity and permeability.

The depth in a sediment column at which lithification of calcareous material begins is not so well indicated as in clay shale. Calcium carbonate is easily cemented and replaced by other minerals, without great pressures; this is evidenced by the formation of beach limestone and by the occurrence of the high-porosity, deep-sea foraminiferal limestones such as samples 1 and 2 of the present study. These two samples came from the tops of two seamounts and presumably had only a light cover of material on top of them. The conclusion here

is that the depth to lithification in a calcareous deposit could be relatively shallow. The factors involved, however, should normally be the same as those for clay shale: pressure, age, and chemical factors are the most important. If calcium carbonate is relatively easy to convert to rock, then one would expect that the depth to lithification in a calcareous section would certainly be no greater (and might be less) than that in a clay-shale section. This would place the limits on the thickness of the first layer of unlithified calcareous sediment, in general, at about the same depths as in a clay-shale section—about 150–700 meters. Areas of faster accumulation might have greater thicknesses of unlithified sediment, inasmuch as a definite relationship exists between age and lithification (Weatherby and Faust, 1935, p. 8; Faust, 1951); older sediments have a longer time to be altered by chemical effects.

After a calcareous ooze has been lithified into a limestone it has a strong, rigid structure, and further deposition causing increased overburden pressures probably would not cause a marked reduction in porosity. Porosities in limestones should decrease because of chemical changes, but the marked reduction of porosity with depth, so well known in shale, will not occur to a significant degree. This is borne out by depth-porosity diagrams in limestone from oil-well data (Wyllie *et al.*, 1956; Hicks and Berry, 1956). This factor allows relatively low porosities in rock near the surface not caused by pressure-induced consolidation. The table and curves illustrating consolidation of calcareous sediments are, therefore, merely suggestive and do not illustrate expectable porosities at depth for lithified calcareous material. The transition between cemented and uncemented calcareous material is likely to be sharp and not gradational.

Limestone is known in mining geology as the host rock most likely to be metasomatically replaced by other minerals; thus replacement phenomena such as silicification, phosphatization, and dolomitization might be expected in a limestone bed.

Foraminiferal limestone.—In the deep sea, including seamounts but excluding the coral-atoll environment with its coralline algae and coral, known occurrences of limestone are rare. Murray and Lee (1909, p. 22) record three instances (one each by the ALBATROSS, CHALLENGER, and BRITANNIA) in which hardened *Globigerina* ooze was dredged from the sea floor. The Scripps Institution-U. S. Navy Electronics Laboratory Expedition in 1950 dredged indurated calcareous ooze in the form of foraminiferal limestone from four seamounts in the Mid-Pacific Mountains west of Hawaii and from one seamount in the Northern Marshall Islands. The geology and paleontology of this material is described in detail elsewhere (Hamilton, 1953; 1956b; Hamilton and Rex, 1959) but is reviewed briefly here. For this study five samples of this material have been examined in great detail, including determinations of the velocity of the compressional elastic wave. These samples are very important in considering the possibility that normal sedimentation and lithification can explain, in part, the present concepts of sea-floor structure based on seismic studies.

Sample 1 was dredged from the top center of Hess Guyot, a peak on the main ridge of the Mid-Pacific Mountains, at a depth of about 915 fathoms (Mid-Pacific Station 33 C; Lat. 17° 49′ N., Long. 174° 17′ W.). The dredge brought up large amounts of rounded, soft but indurated foraminiferal limestone in the form of rocks ranging from cobbles to boulders a foot in diameter. The material was so soft when wet that the best explanation for the rounding was that it had taken place during the mile-long pull back to the sea surface. This may have been the surface material, but possibly the heavy steel dredge could have hooked chunks of it from an indurated expanse under unlithified sediment; small pipe dredges, attached to the large dredge, were filled with Recent planktonic deposition.

The foraminiferal limestone of sample 1 was dated as late Paleocene (Hamilton, 1953) by its uncontaminated planktonic foraminiferal fauna. This material was examined under the petrographic microscope and by X-ray diffraction and determined to be composed of planktonic Foraminifera and finely divided calcareous material cemented together by calcium carbonate. Other physical properties of this material are indicated in Table 3.

Sample 2 was dredged from the break-in-slope on the top of Guyot 19171, a flat-topped seamount in the Mid-Pacific Mountains, at a depth of about 750 fathoms (Mid-Pacific Station 26 A-3; location restricted). The dredge haul brought up slabs and boulders of olivine basalt and one boulder of foraminiferal limestone (14½ x 14 x 6 inches) covered by a 6-mm coat of ferromanganese.

The limestone of sample 2 (Pl. 1, fig. 1) was dated as lower Eocene by uncontaminated planktonic foraminiferal fauna (Hamilton, 1953). The original calcium carbonate ha

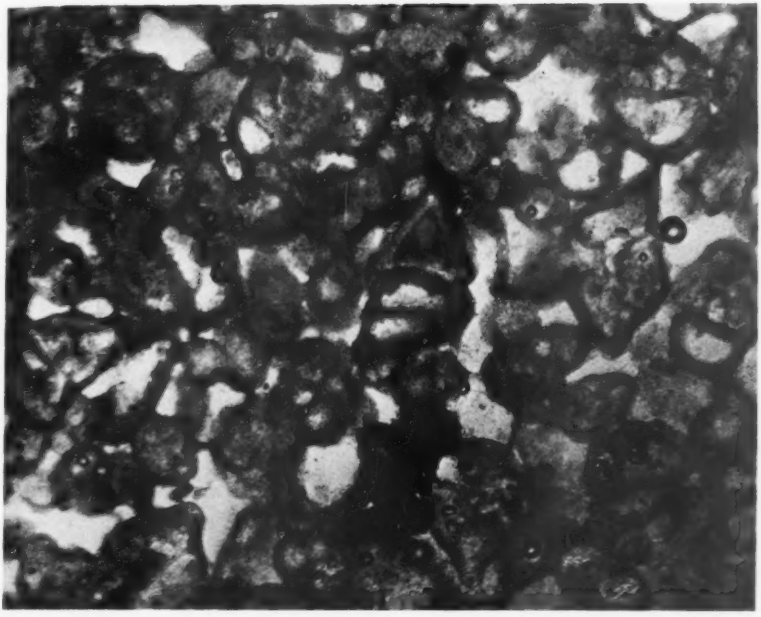

FIGURE 1. — Sample 2, Mid-Pacific Station 26-A3

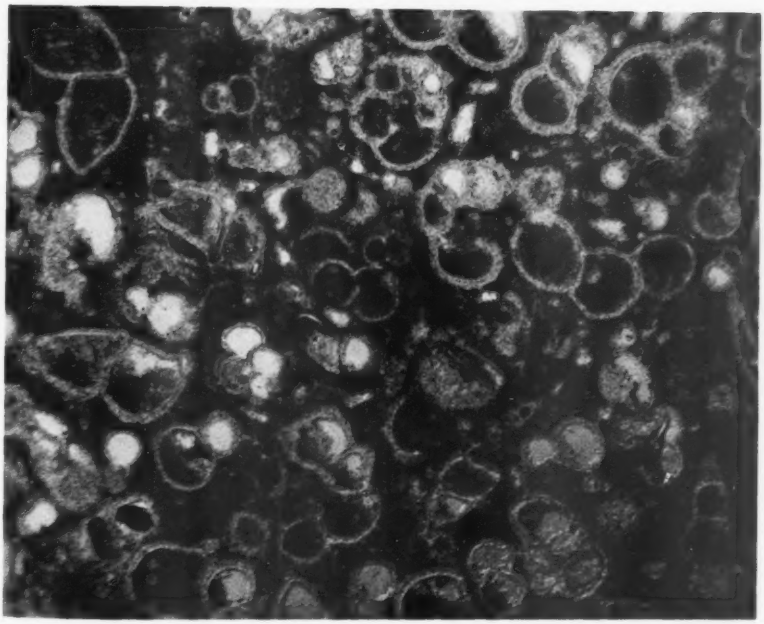

FIGURE 2. — Sample 5, Mid-Pacific Station 37 C

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been partially replaced by phosphate as shown by petrographic study and by X-ray-diffraction studies by Rex (Hamilton and Rex, 1959). This

TABLE 3.—PROPERTIES OF LITHIFIED *Globigerina* OOZE

Sample no.	Density solids (gr/cc)	Porosity (per cent	Velocity (km/sec)*
1	2.70	68	1.77
2	2.87	62	1.95
3	2.83	24	3.86
4	2.86	12	4.75
5	2.85	5	5.46

* ±0.02

limestone was analyzed spectrographically by C. E. Harvey of the American Spectrographic Laboratory with the following results:

Elements as oxides	*Per cent*
Calcium CO_2 and anions	65.5
P_2O_5	28.0
SiO_2	1.9
Other	4.6

Other physical properties of this material are indicated in Table 3.

Samples 3, 4, and 5 (Pl. 1, fig. 2) were dredged from the break-in-slope on top of Cape Johnson Guyot, a peak on the main ridge of the Mid-Pacific Mountains, at a depth of approximately 1090 fathoms (Mid-Pacific Station 37 C; Lat. 17° 10′ N., Long. 177° 10′ W.). The dredge brought up about 200 pounds of highly fossiliferous material containing shallow-water rudistids and corals from the Middle Cretaceous imbedded in foraminiferal limestone which was dated by its contained Foraminifera as probably late Paleocene; this date determination for the foraminiferal limestone was made recently and is reported here for the first time. Thin-section study, X-ray diffraction, and spectrographic analysis show that this material is a foraminiferal ooze in which calcium carbonate has been partially replaced by phosphate (francolite or dehrnite; Hamilton and Rex, 1959). Spectrographic analysis showed the following:

Elements as oxides	*Per cent*
Calcium CO_2 and anions	61.2
P_2O_5	35.0
SiO_2	1.2
Other	2.6

Other physical properties are listed in Table 3.

The velocities of compressional elastic waves were determined in samples 1–5 through the courtesy of the California Research Corporation, La Habra, California. Velocity through the samples was measured by the conventional method of timing an elastic pulse generated by a quartz transducer through a sample of known length. All determinations were made at room temperature and 1 atmosphere of pressure; the frequency was about 500 kc; the pulse length was 2.1 microseconds. All samples were saturated: sample 1 with silicone oil, sample 2 with sea water, and samples 3, 4, and 5 with distilled water. Saturation of 1 and 2 was done under pressure and of 3, 4, and 5 by boiling in a vacuum. Velocities ranged from 1.8 to 5.5 km/sec. Velocity, density of solids, and porosity data are summarized in Table 3.

COMPRESSIONAL WAVE VELOCITIES IN DEEP-SEA SEDIMENTS AND ROCKS

Velocities in Unlithified Sediments

Until about 3 years ago the only velocity determinations for "unconsolidated" sea-floor sediments were a few rough (thick-section) determinations by seismic work at sea. Readable returns are not usually obtained from seismic refractions in the surface sediments of the sea floor; as a rule a seismic velocity is assumed for this upper layer. The assumed velocity is usually an average based on data available from other studies.

The principal properties of unlithified, saturated sediments that affect sound velocity are the grain size, shape, and mineral composition, properties of the interstitial water, sediment structure (which includes intergrain attractive forces), and porosity. Sound velocity will be ultimately controlled by the joint effect of these various properties on the elastic structure of a sediment. Up to now no valid way has been determined to measure, directly, the elastic properties of sediments; therefore one must seek substitute properties that reflect, or are related to, elastic properties. The most easily derived of such properties appears to be porosity because porosity is a measure of the openness of the aggregate of minerals that makes up the sediment and consequently is likely to be an indication of the rigidity of the sediment structure (Hamilton *et al.*, 1956, p. 14). The reader further interested in the factors affecting velocity should read the recent sum-

maries by Hamilton *et al.* (1956), Shumway (1956; 1958; 1958, Ph.D. thesis, Univ. California), Laughton (1954; 1957), Nafe and Drake (1957), and Sutton *et. al.* (1957). Some of these data are summarized in Figure 7.

Recent papers on seismic work at sea (Hill, 1957; Officer, 1955; Officer *et. al.*, 1957; M. Ewing, *et al.*, 1955; J. Ewing *et al.*, 1957; Katz and Ewing, 1956; R. L. Fisher, 1957, Ph.D. thesis, Univ. California; Nafe and Drake, 1957; Gaskell, Hill, and Swallow, 1958) indicate that seismic velocity in unlithified marine sediments is largely confined to the range 1.5–2.2 km/sec. These velocities are averages for an uncertain thickness of sediments and uncertain sediment types, and velocities of more than 2.0 km/sec probably indicate some degree of lithification. Laboratory measurements of samples from the deep sea by Laughton (1954), Shumway (1956; 1958; 1958, Ph.D. thesis, Univ. California) and Sutton *et al.* (1957) indicate the following values at the sediment surface for red clay and calcareous ooze at about room temperature and 1 atmosphere of pressure:

Sediment type	Velocities, km/sec		
	Shumway	Sutton	Laughton
Red clay*	1.48–1.57	1.43–1.57	1.47–1.54
Calcareous ooze	1.49–1.52	1.48–2.77†	1.54–1.63

* Red clay defined as a deep-sea clay with less than 30 per cent calcium carbonate

† Sutton *et al.* think that their higher values (2.37 and 2.77) are probably due to the rigidity imparted by the first stages of lithification.

Shumway (1958) showed that correction of the velocity from room temperature to *in situ* values should usually be on the order of −0.08 km/sec. Laughton (1957) showed that the effect of hydrostatic pressure is the same in sediment as in water; as increase of about 3–4 per cent would be expected from values at 1 atmosphere pressure to values at most sea depths.

Most geophysicists doing seismic work at sea agree that a positive velocity gradient downward exists in sea-floor sediments although in many cases an average figure is taken, and the seismic model adopted to analyze the data assumes no gradient. The laboratory work of Laughton (1954) and seismic work previously cited indicate that in deep-sea sediment one should expect a positive velocity gradient of 0.5–3.0 km/sec^{-1}.

An interesting and important phenomenon is the relationship between the velocity of sound in the sea water just above the bottom and the velocity in the surficial clay; frequently the velocity is less in the sediment. Hersey *et. al.* (1952) and Officer (1955) first illustrated the phenomena for seismic work at sea; Katz and Ewing (1956) further discussed the relationship from their, and other's, seismic work and concluded that, where present, the depth of low velocity in the sediment may extend to 50 m below the sediment interface. Hamilton *et al.* (1956), Shumway (1956; 1958; 1958, Ph.D. thesis, Univ. California), and Sutton *et al.* (1957) noted and discussed the low-velocity relationship in the laboratory and in *in situ* shallow-water measurements. A velocity of sound less in sediment than in the bottom water is likely to be present in clays and silts where the porosity is greater than about 60 per cent.

Velocities in Lithified Sediments

General.—The general subject of seismic velocities in lithified sediments (sedimentary rocks) is of great importance in seismic exploratory work in the oil industry, and lately, because of seismic work at sea, the subject has assumed critical importance in our understanding of the structure and history of the ocean basins. Considerable laboratory and field work has been done to determine the velocity of sound in sedimentary rocks. Krumbein summarized the subject in 1951, and recent papers by Hughes and his co-workers (1950–1952), Wyllie, Gregory, and Gardner (1956), and Nafe and Drake (1957) summarize the factors and values to date.

The properties of sedimentary rocks that directly affect sound velocity are, of course, the elastic properties. Mineral grain size, composition, orientation and shape, pressure, temperature, pore liquids, degree and kind of cementation, density, and porosity are all influential factors. As in the unlithified sediments, sound velocity will be ultimately controlled by the joint effect of these various properties on the elastic structure of the rock. When a sediment is lithified by cementation of the grains, a material that has a rigid structure is produced. Many investigators have determined and studied the elastic properties of various rocks; this cannot be done by direct measurement on unlithified sediments. For example, Weatherby and Faust (1935) and Horton (1943) show how the rigidity of a lithified sedi

ment increases faster than the density during the porosity reduction because of gravitational consolidation and lithification.

The velocity of sound in rocks increases with the age of the rocks. This increase of velocity appears to be linked with the general increase of lithification of sediment with age (Weatherby and Faust, 1935, p. 8; Faust, 1951).

The importance of porosity in velocity in rocks is the same as that previously discussed for unlithified sediments: it is a rough measure of the openness of the rock structure and therefore a relatively easily measured physical property which is related to the elastic properties such as rigidity. Wyllie, Gregory, and Gardner (1956) have shown that porosity variations considerably outweigh the influences of differing materials, pressures, and other environmental factors. Thery show (p. 65) an "Experimental Average Curve" (Fig. 7) of velocity plotted against porosity for limestone and sandstone. They then show that the curve (representing, of course, an elongate area of data spread) is valid under different conditions in laboratory and field, including oil wells where the estimated velocity from the porosity data compares favorably with the velocity as determined by the new continuous velocity-logging technique.

The scatter of data along any porosity-velocity curve must represent the effect of factors other than porosity. One of the most important of these is the material forming the solid matrix of the rock. Wyllie, Gregory, and Gardner (1956, p. 45; personal communication from Wyllie) show that the velocity of sound in quartz and calcite varies according to the crystallographic axis; a random orientation of axes in a calcareous rock would yield a frame, or matrix, velocity of 22,300 feet per second (6.8 km/sec); a random orientation of quartz grains would yield a frame velocity of about 19,500 feet per second (5.9 km/sec). This latter figure is confirmed by the high-pressure-velocity studies of Hughes and Jones (1951, p. 176, Fig. 7) who reported that a sample of sandstone reached a velocity of about 5.7 km/sec when it was crushed to "zero" porosity. Wyllie *et al.* use an empirical frame velocity of 17,500 feet per second (5.3 km/sec) for shale. These frame velocities probably represent the highest possible values for rocks at zero porosity, a condition rarely attained.

Velocities in shale.—Faust (1951) made a study of the variations of velocity with depth and geologic time for a large number of dominantly shale sections. He utilized actual velocities from more than 70 wells on the North American continent to form curves for Eocene and Tertiary (post Eocene) (Fig. 9). In these curves the velocity increases from about 2.1

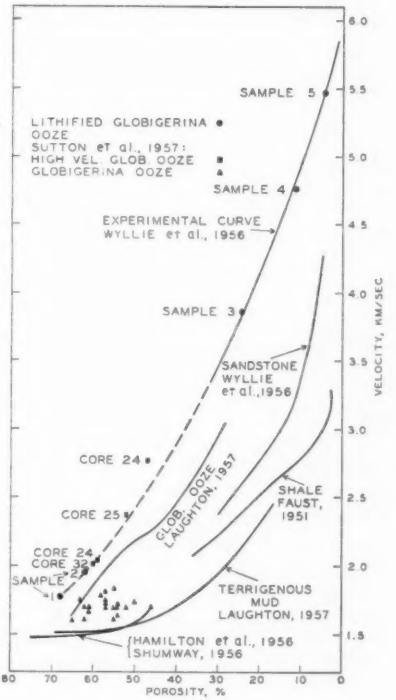

FIGURE 7.—POROSITY VS VELOCITY OF THE ELASTIC COMPRESSIONAL WAVE IN VARIOUS SEDIMENTS AND ROCKS

km/sec at 300 m to 3.32 km/sec at 3000 m. These curves, because of the great amount of data involved, can be taken as expectable variations of velocity with depth in Tertiary shale sections (which may include minor sandstone layers). The curve for Tertiary (post-Eocene) shale is correlated with the porosity information of Skeels (*in* LeRoy, 1950) and shown as a typical curve for porosity velocity in Figure 7.

Faust (1951) showed an increase in velocity caused by lithification because of age (the longer in place the better chance for more cementation); probably the highest velocities expectable in shale are indicated in his Pennsylvanian shale (p. 194) which reaches velocities of about 3.9 km/sec at 2000 m and 4.4 km/sec at a depth of 3500 m; in some sections Faust records velocities to 4.5 km/sec.

Many writers have shown the relationship between depth and increasing velocity in shales. Wyllie *et al.* (1956) show that this is largely due to changes of porosity with depth (as they reflect the increasing rigidity of the rock structure). When a shale is metamorphosed into a slate the value of velocity along the "bedding" plane may be as high as 6.3 km/sec.

Shepard and Moore (1955 p. 1587) note that shalelike partings occur in some of their Gulf of Mexico sediments at depths of only 69 feet. Laughton (1957, p. 249) has shown a definite anisotropy developed during laboratory compaction of a clay; the velocities along the "bedding plane", or normal to the pressure direction, are 3–5 per cent higher. This anisotropy parallel to the bedding planes is well known in geophysical prospecting in stratified rocks. Velocities parallel to the bedding planes in shale may be as much as 40–50 per cent higher than those normal to the bedding planes (LeRoy, 1950, p. 1089; Uhrig and Van Melle, 1955). LeRoy (1950, p. 1089) thought that an average figure for this difference would probably be about 10 per cent. This anisotropic effect is present, to a lesser degree in carbonate sediments. These facts indicate that, in shale, transverse anisotropy is probably a pressure effect of the overburden. The anisotropic effect should be considered in oceanic areas where stratified sediments possibly occur.

In summary: the velocity of the elastic compressional wave in shale sections will vary according to the porosity and depth and will be higher parallel to bedding planes. The minimum and maximum velocity values normally expectable to a depth of 2 km will be from about 1.8 to about 3.9 km/sec; probably most noncalcareous, post-Cretaceous shale sections will vary within narrower limits with a maximum, normally expectable velocity of about 2.6 km/sec at a depth of about 2.0 km and about 2.1 km/sec at 300 m.

Velocities in limestone.—The work of Wyllie, Gregory, and Gardner (1956) agrees very well with, and extends, previous work on limestone both in field and in the laboratory; their "Experimental Average Curve" (p. 65) can be considered as a summary of expectable velocities in limestone.

The validity of Wyllie, Gregory, and Gardner's curve was demonstrated in the present study when the velocities of the compressional waves in samples 1–5 were very closely predicted. The final plot of the sample velocities vs porosity lie extremely close to the "Experimental Average Curve" (Fig. 7); the closeness to the curve was, of course, a statistical coincidence in that the curve represents an area of data.

As pointed out by Hicks and Berry (1956, p. 753) and Wyllie *et al.* (1956; 1958), one must be careful in transferring studies of pressure vs velocity from the laboratory to natural conditions in the field. Many laboratory studies involve specimens jacketed by some material: when the specimen is placed under pressure most of the pressure may be transmitted to the pore water and relatively little to the rock structure. Conversely if the specimen is not completely saturated the pressure will be borne by the rock structure alone, and usually the velocity will be markedly higher.

A saturated, homogeneous, porous rock at depth below the sea floor will have the same pressure system as a sediment: a pore-water pressure and an intergranular pressure. Wyllie *et al.* (1956; 1958) have shown that the effect of high pressure in such rocks is to close any flaws or cracks and cause the velocity to be approximately related to porosity in a simple "time-average" relationship:

$$\frac{1}{V_m} = \frac{n}{V_w} + \frac{1-n}{V_{sol}}$$

where V_m = measured velocity, or velocity in the saturated rock; V_w = velocity in the pore water; V_{sol} = velocity in the matrix, or solids; n = porosity.

This "time-average" relationship is very close (±3 per cent in velocity for limestone) to the "Experimental Average Curve" (Fig. 7). The conclusion here is that, in limestone, the porosity is the most important easily measured physical property (including pressure) that has an apparent relationship to velocity. The velocities expectable in a limestone at any depth, therefore, can be estimated accurately from its porosity and will vary approximately along the line of the "Experimental Average Curve" regardless of pressure.

Sample 1 of the present study is a foraminiferal limestone with a measured velocity of 1.5 km/sec and a porosity of 68 per cent. This specimen probably represents a near minimum value of velocity for limestones, although lower values are theoretically possible. Randomly oriented calcite grains in a pure calcareous rock of zero porosity would have a velocity of about 6.8 km/sec, which represents the highest theoretical value for velocity in a limestone. Field and laboratory data, however, show that the highest value normally to be expected should

be about 19,500 feet per second (5.9 km/sec; Wyllie *et al.*, 1956, p. 62–65), although a few velocities of the range 6.1–6.45 km/sec have been reported (Birch *et al.*, 1942; Hicks and Berry, 1956, p. 753).

Velocities in mixed materials.—Inasmuch as shale velocities may be as high as 4.5 km/sec, and limestone velocities overlap this range from about 1.8 km/sec, a material with velocities less than 4.5 km/sec cannot be identified on the basis of velocity alone. Sutton *et al.* (1957) have shown that the velocity of sound increases in a sediment with increasing calcium carbonate content. One would expect that calcareous clays would have velocity values intermediate between clay and calcareous ooze. Another complicating factor is that velocities in volcanic material range from values as low as 4.0 km/sec to the values for the supposed basalt of the earth's outer crust (approximately 6.8 km/sec; Raitt, 1956). One of the purposes of this paper is to show that normal sedimentation and probable lithification might explain some of the higher velocities found in many places below the unlithified sediments (the second layer) in the deep seas. A layer of volcanic rocks could also explain these higher velocities, but the idea that all the ocean basins have a layer of basalt or volcanic rocks at about the same depths within the sediments is the less likely alternative. However, in many areas, especially near islands, the higher velocities undoubtedly represent volcanic rocks.

If the "time-average" formula is used for approximating the velocity in a rock, a matrix velocity of 22,300 feet per second should be used for limestone and 17,500 feet per second for shale; the velocity of sound in the bottom water of the north-central Pacific at 5-km depth is about 5060 feet per second (E. R. Anderson, personal communication). Approximate values for mixtures of rock types can be computed by combinations of these values.

Discussion

Figure 7 is a summary of recent information on the relationship between velocity and porosity in sediments and rocks; it differs from other, similar, figures in that it differentiates sharply between various types of sediments and rocks. A universal equation or relationship between porosity and velocity can never be worked out, because sediments and rocks have entirely different structures with entirely different elastic properties. Even in unlithified sediments no valid universal equation is possible, as is readily apparent by considering the porosity-velocity data for silts and clays versus that for calcareous sediments. Sutton *et al.* (1957) have shown that velocity increases with increasing percentages of calcium carbonate. At least two reasons for this are readily apparent: (1) calcite has a higher velocity than quartz or clay particles, and (2) the grain-to-grain structure of a calcareous "sand" is more conducive to high velocities than a silty clay of the same porosity with its grains separated by adsorbed water layers.

The writer has extended the "Experimental Curve" of Wyllie *et al.* along the dashed line from this curve in Figure 7, so that the two high-porosity samples (1 and 2) could be included; this curve is a median for Wyllie's data so that any such extension is merely suggestive. Unlithified calcareous sediments should lie along the low-velocity side of this curve (as they do). A rough measure of the increase in velocity due to lithification (without changing the porosity) is indicated by these curves. For example: if a calcareous ooze similar to that of Laughton's (1957) with 30 per cent porosity were converted to a limestone with 30 per cent porosity an increase of velocity on the order of 1700 feet per second should occur. The fact that the anomalously high velocities of Sutton *et al.*'s cores 24, 25, and 32 fall on or near the extension of Wyllie *et al.*'s curve is probably significant and indicates that Sutton *et al.*'s supposition that these sediments had started to lithify is correct.

Nafe and Drake (1957, p. 525–526) noted that when their seismic data for velocity vs depth are plotted with Laughton's curve from studies of artificial compaction of calcareous ooze, Laughton's curve forms a lower boundary for their data; thus all their sediments have higher velocities than Laughton's at the same depth. The big problem in relating laboratory consolidation studies to field relationships is that of time. Primary laboratory consolidation occurs in a matter of hours, or days; deep-sea sediments are consolidated in thousands and millions of years. This means that an "unconsolidated" sediment in an ocean basin has plenty of time to form strong intermolecular bonds as well as to progress in incipient lithification because of chemical effects. A sediment in place should, therefore, always have a higher velocity than the same sediment, removed, disturbed, and artificially consolidated.

Conclusions

Expectable Consolidation and Lithification in an Area of Clay Deposition

In an ancient ocean basin whose depth, temperature of bottom waters, and other factors have caused continuous deposition of clay, the following sequence of events should normally be expected:

The slowly deposited clay follows the expectable consolidation for clay previously discussed. As the intergranular pressures build up, and with pore-water excess pressure equal to zero (the hydrostatic case), the material consolidates along the lines shown in Table 1 and Figure 2. In the slowly consolidating sediment decrease in porosity to pressures around 50 kg/cm² is relatively fast because of mechanical rearrangement of grains and expulsion of most of the free pore water and much of the adsorbed water around the grains. When the porosity is between 35 and 45 per cent contact between grains should start lithification of the sediment and mechanical deformation of grains, solution, recrystallization, and other chemical changes. In short, at porosities between 35 and 45 per cent, the sediment is no longer a loose aggregate but may have started to lithify and become a rock. It is significant that the upper porosities of shale are in the 35–45 per cent range.

Table 1 and Figure 4 show that a porosity between 35 and 45 per cent is reached at depths between 150 and 700 m (depending on chemical factors and grain densities); Figure 4 also indicates that the probable average depth to lithification is nearer 300–400 m.

People familiar with seismic research at sea will immediately recognize that almost everywhere a top layer of low-velocity material lies, on the average, within the limits noted above. The work of Hill (1952; 1957), Hersey *et al.* (1952), Officer (1955), Katz and Ewing (1956), Nafe and Drake (1957), and others indicates that there is a fast changing positive-velocity gradient in deep-sea sediments to around 300 m, below which the gradient is less. Nafe and Drake (1957), in a special study, noted the common occurrence of refraction and reflection horizons at depths from 400 to 800 m in Atlantic sediments. In a recent excellent paper concerning seismic exploration in the Atlantic and adjacent seas, Ewing and Ewing (1959) noted transformation to shear waves and refraction and reflection horizons which they presume may have taken place from semiconsolidated layers within the sediment column. In the Arctic Basin, Crary and Goldstein (1957) noted reflection arrivals from depths of 250–300 m, where the average interval velocities are 2.04 km/sec. In the Pacific, in the northeast red-clay area, Raitt (1956) noted an average first-layer thickness of 290 m; in the east-central Pacific calcareous area the thickness is 330 m. Shor (1959) showed a reflection horizon in the same area at depths between 200 and 300 m. Gaskell, Hill, and Swallow (1958) showed, in all oceans investigated, the presence, or probability, of a topmost layer of low velocity. In many areas there are thicker sections of presumed unlithified material of 1 km or more. In all but a few of these determinations the writers do not exclude the possibility of masked layers in the sediment column, and further exploration may well show shallower reflecting horizons.

The presence, in all the ocean basins, of a topmost layer on the order of 300 m in thickness, can hardly be the result of chance. It is important evidence on the possible depths to lithification previously discussed, and is probably the expression of a universally valid phenomenon.

Throughout geologic time a sediment column, in an area of slow deposition and consolidation would have, at the top of the column, about the same thickness of unlithified material. This thickness will ordinarily be on the order of 150–700 m. A natural consequence of this would be that the lithified portion of the column (the second layer) would gradually increase in thickness while the topmost portion remained fairly constant in thickness. This postulate (and prediction) of normal consolidation and lithification is illustrated in Figure 8 which is a seismic section across the deep basin east of the Marquesas and up onto the Southeastern Pacific Plateau. This single environment of calcareous deposition has a first-layer average thickness of 240 m (range in five stations: 200–280 m), but the second layer of presumed lithified material ranges from 730 m to 1590 m.

If the above hypotheses on lithification are correct what would be the normally expectable velocities at depth in a post-Cretaceous, noncalcareous clay-shale section? Study of the laboratory work of Laughton (1954; 1957), Shumway (1956; 1958; 1958, Ph.D. thesis, Univ. California), and Sutton *et al.* (1957) indicates that the sediment surface should have a velocity of about 1.5 km/sec. If a conversion of clay to shale at 300 m is assumed (available data indicate that this is a reasonable median figure) then a study of Figure 9 indicates that at 300 m, the velocity should be about 2.1 km/sec; this would yield an average velocity in the first 300 m of about 1.8 km/sec (although the

gradient is not a straight line). Inasmuch as the velocity increases with depth in the lithified portion of the section, the average velocity in this section will be dependent on its thickness.

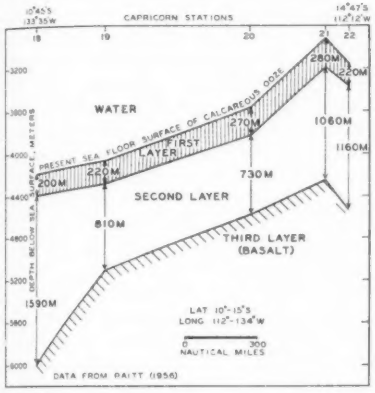

FIGURE 8.—VELOCITIES AND THICKNESSES OF LAYERS IN THE SEA FLOOR OF CALCAREOUS AREA, EAST-CENTRAL PACIFIC
(From Raitt, 1956)

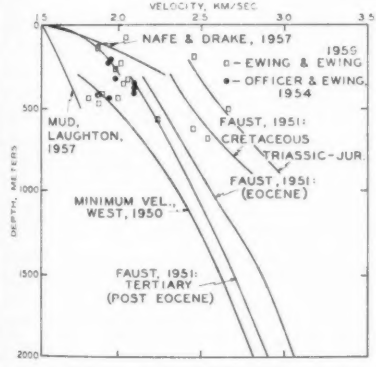

FIGURE 9.—VARIATIONS OF VELOCITY WITH DEPTH

Shale sections (Faust, 1951; West, 1950) and unlithified sea-floor sediments to 300 m (Nafe and Drake, 1957) compared with seismic-refraction information of Officer and Ewing (1954) and Ewing and Ewing (1959). These field determinations are contrasted with laboratory experiments on terrigenous mud (Laughton, 1957).

With the above hypotheses in mind one can predict the results of seismic exploration in a thick section of noncalcareous post-Cretaceous silt-clay material as follows:

(1) An upper, relatively unlithified section will range in thickness from 150 to 700 m and average about 300–400 m; in this section the velocity should be about 1.5 km/sec at the top and 2.1 km/sec at the bottom; average velocity should be on the order of 1.8 km/sec.

(2) The lower, lithified section of clay shale varies widely in thickness because it is being added to at the top by the upper layer, which maintains a relatively constant thickness. In this lower layer the average velocity will vary according to the thickness of the layer about as follows (derived from Figure 9 by taking an average between the Eocene and post-Eocene lines):

Thickness of second layer (m)	Velocity at top (km/sec)	Velocity at bottom (km/sec)	Average velocity (km/sec)
500	2.10	2.41	2.25
1000	2.10	2.58	2.34

Most velocity figures for the second layer that appear in the literature are valid for only the top of that layer.

Any calcareous material will cause the velocity to be higher; these estimates are for a clay-shale section only.

An almost ideal realization of the above predictions is shown in Officer and Ewing (1954). In two sections across the continental shelf and slope, and into the deep-sea basins, they record increasingly thick sections of "unconsolidated sediments." We see in these sections a topmost section ranging in thickness from 60 to 580 m (average of 343 m) and with an average velocity of 1.73 km/sec; the second layer ranges in thickness from 40 to 2870 m.

In the recent paper by Ewing and Ewing (1959), which was published after the above predictions were written and submitted for publication, there are several areas that possibly illustrate these relationships. A summary of these areas follows:

Area	Average first layer		Average second layer		Average total thickness (layers 1 and 2) (km)
	Thickness (km)	Velocity (km/sec)	Thickness (km)	Velocity (km/sec)	
Western Atlantic (Stations G-10, 12, 13)	0.32	1.70	0.88	2.16	1.20
Western Atlantic (Station A-157-29)	0.44	1.62	1.09	1.82	1.53
Greenland Sea (Station F-4)*	0.57	1.79	0.95	2.23	1.52
Norwegian Basin* (Stations F-7, 9)	0.29	1.75	0.70	1.94	0.99

* Higher-velocity layers are present below these two layers (Ewing and Ewing, 1959, p. 296, 311).

Ewing and Ewing (1959) believe that the second layer is semiconsolidated sediment; the term "semiconsolidated" was used by these writers to indicate partially lithified material.

The velocities at the tops of the second layers determined by Officer and Ewing (1954) and Ewing and Ewing (1959), discussed above, are plotted in Figure 9. They are close to the indicated curves for shale, much closer than to the velocities expected at depth in an unlithified section of terrigenous mud as shown by Laughton's curve (Fig. 9) which has been converted from pressure to depth.

An excellent paper by Officer (1955) concerning a reflection profile in the Atlantic abyssal plain at a water depth of 5.1 km also illustrates the above concepts. Officer found the velocity in the upper 100–150 feet of sediment to be lower than that in the bottom water; a prominent reflector occurred at a depth of 357 m, and Officer stated (p. 275) that it was present on all deep-sea profiles examined to date. The average velocity in this upper 357 m was 1.7 km/sec. The lower part of the column was 663 m thick and had an average velocity of 2.2 km/sec.

Expectable Consolidation and Lithification in an area of Calcareous-Ooze Deposition

Calcareous-ooze deposits should consolidate along the lines indicated in Table 2 and Figure 2. A relatively rapid decrease of porosity under accumulating intergranular pressures to values around 60 per cent should occur where the pressure is about 19 kg/cm². At greater pressures the porosity decrease is much slower than in a clay deposit. During this consolidation, after pressure has caused a grain-to-grain contact, pressure-point solution, reprecipitation, and other chemical effects should cause the ooze to lithify. This cementation and recrystallization would continue after consolidation and result in further reduction of porosity. After lithification, however, a limestone does not continue to decrease in porosity because of increased overburden pressures as is the case in shale sections.

The more important factors governing depth to lithification in a calcareous section should normally be the same as those involved in lithification of clay into shale: pressure, chemical factors, and age. Inasmuch as calcareous material is more readily altered than most other sedimentary materials, the depths to lithification in a calcareous section should be about the same as in a clay section (150–700 m) or perhaps less. The thickness of the first layer in the Northeast Pacific red-clay area averages 290 m (range 170–450 m); in the East Pacific calcareous area south of the equator the first layer averages 260 m (range 200–340 m); in the Equatorial area of maximum accumulation of calcareous material (Arrhenius, 1952) three stations indicate a first-layer average thickness of 490 m (range 380–570 m). All these thicknesses are within the previously discussed approximate general range of 150–700 m; the increased thickness under the Equatorial current areas thus require no special explanation although increased rates of deposition are probably a factor, because age is a factor in degree of lithification (Faust, 1951).

The discovery that foraminiferal limestone derived solely from the lithification of calcareous ooze has velocities ranging from 1.8 to 5.5 km/sec indicates that many of the velocity values determined by seismic work at sea might be explained by normal deposition, consolidation, and lithification. The literature on velocities of sound in rock indicates further, as previously discussed, that values between 5.5 and 6.5 km/sec also can be explained by low-porosity beds of these normal deep-sea deposits of calcareous ooze that have been lithified to form low-porosity limestone beds.

Composite Sections of Clay and Calcareous Sediments

If the highest velocity expectable in a shale section is approximately 4.5 km/sec then one has to explain the high velocities found in many places under some present areas of red clay. These velocities range from 4.5 to 6.04 km/sec in the northeast Pacific Basin (Raitt, 1956). The simplest supposition here is that the lower layer is composed of foraminiferal limestone lithified from oozes deposited during a former time of calcareous deposition in the area.

The deposition of calcareous ooze in the deep sea probably began with the evolution of the planktonic Foraminifera in the Late Mesozoic although earlier deposition of coccolith-rich sediments is possible (Bramlette, 1958). The paleontologic record on land has indicated this blossoming of planktonic faunas for some time, but only in the last 10 years has sampling shown that ancient deposits were truly present in the deep-sea areas. Upper Cretaceous planktonic material was cored at the base of a flat-topped seamount in the Central Pacific, and pure, uncontaminated lithified and unlithified Paleocene and Eocene deposits of *Globigerina* ooze were

dredged and cored on the tops of several of the flat-topped seamounts (Hamilton, 1956b). Coring in both the Atlantic and Pacific deep-sea areas has shown that Tertiary calcareous material in some places is near or at the surface. We thus know that these planktonic Foraminifera were collecting on the sea floor as pure deposits of calcareous ooze during the late Cretaceous and early Tertiary. The argument for a wider distribution of calcareous ooze during the Late Cretaceous and Early Tertiary than today has been discussed by a number of writers, as noted below.

Flint (1957, p. 488) summarized the data on the cooling climates on land and thought that the evidence showed a secular cooling that had been under way since the Eocene. This is in agreement with Durham (1950, p. 1252–1260) who thought it probable that the 18.5° C. isotherm in the sea was in the vicinity of Lat. 53° N. during the Late Cretaceous. During the Paleocene and Eocene the 20° C. isotherm was probably north of Lat. 49° N.; from the Eocene to the Pliocene it shifted southward, possibly with minor fluctuations, to its present position. Studies by Emiliani (1954) of the temperature of Pacific bottom waters and the probable temperature of polar waters during the Tertiary are in agreement with the evidence from the continents. In brief, during the Late Cretaceous and Early Tertiary, tropic and semitropic conditions existed far to the north and south of their present position, polar waters were much warmer than today, and consequently the bottom waters of the oceans, which depend on sinking polar waters for their basic temperature, were much warmer than at the present time. A slow cooling of the bottom waters during the Tertiary accompanied a similar cooling of the polar water as the semitropic climates moved back toward the equator.

The factors causing deposition and accumulation of calcareous ooze are not yet completely understood, but two of the primary factors are depth and temperature of the bottom water; colder water, rich in CO_2, is more likely to dissolve the calcareous material on the bottom. These problems were thoroughly discussed by Revelle and Fairbridge (1958). If Late Cretaceous and early Tertiary bottom water was warmer and perhaps slightly shallower, the areal extent of calcareous ooze deposition was probably much greater than today.

Daly (1944, p. 1388–1389) postulated that a layer of calcareous material 500–600 meters thick lay over most of the area of present volcanic islands in the Pacific and that this explained the presence of the necessary calcium carbonate, which, when mixed with the basalt magma of the incipient islands, yielded the proper chemical constituents to produce melilite and nepheline basalt. In discussing this idea, Macdonald (1949, p. 1585) agreed that the evidence was good for a wider distribution of calcareous ooze in the time before the islands broke through the sediment surface during the Tertiary; he thought that limestone assimilation was not necessary to the production of the high calcium rocks, but that it may have taken place and, if so, furnishes a ready explanation for the high-calcium rocks.

Direct evidence of the wider latitudinal spread of calcareous deposits in the Tertiary was obtained by Arrhenius (1952, p. 195) who made a thorough study of the transitional area from calcareous ooze to the clay of the northeastern Pacific at about Lat. 8° N. He found that in the Tertiary the compensation surface (ooze-clay boundary) dips to the north and is older the farther north it is found.

In summary, then, if during Middle Cretaceous to Tertiary time deposition of planktonic calcareous material began and was much more widespread than today, a bed of calcareous-ooze limestone of high velocity probably underlies much of the north Pacific red-clay areas. This bed should be much thinner in the north than the south to reflect the southward movement of the isotherms as the climate cooled during the Tertiary; it should crop out at the present boundary between clay and ooze deposits and should dip north.

An important aspect of the above situation is that this bed of Cretaceous-Tertiary calcareous ooze should overlie any clay beds deposited in the ocean basins from the formation of the basins to the Late Mesozoic.

What would be the seismic picture in an ancient ocean basin in which clay had been deposited from formation of the basin to the Middle Cretaceous, calcareous ooze from Middle Cretaceous to sometime in the Tertiary, and then, because of cooling of the bottom waters (and possibly greater depths caused by more water volume, isostatic adjustments, and other factors), a return to clay deposition? Figure 10 illustrates a simple seismic model of this situation as it might have occurred in the northeastern Pacific. The early clay deposits have been lithified into a shale, and the calcareous ooze lithified into a deep-sea foraminiferal limestone; above the limestone layer lies unlithified sediment that could be calcareous ooze, clay, or mixtures of the two.

If it were supposed that only red clay lay above the limestone then such a supposition is in question because of the thicknesses of original deposits required and the probable rates of deposition of red clay (Roger Revelle, personal communication): at the highest rate of deposition of red clay (about 0.5 cm/1000 years) any section now more than about 300 m thick would require more than 100 million years of original deposition, (*see* Table 4), which would be before the known advent of the planktonic Foraminifera which probably initiated the extensive deposition of calcium carbonate in the deep oceans.

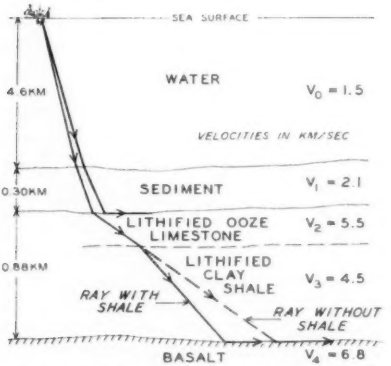

FIGURE 10.—SEISMIC MODEL OF THE DEEP-SEA FLOOR WITH HYPOTHETICAL LAYERED STRUCTURE WITHOUT VELOCITY GRADIENTS

This model illustrates: (1) the second-layer velocity is from the top of the second layer, (2) owing to Snell's Law no refraction occurs in the third layer, and (3) ray paths with and without a third shale layer.

Although calcareous ooze has a slightly higher velocity than clay, this difference is not noticeable nor used in present-day seismic work; a common velocity is usually used in both red-clay and ooze areas; thus all we can say about the top of the second layer is that the velocity could indicate limestone, but this does not mean that the material immediately above the high-velocity material is clay; a layer of unlithified calcareous material could lie above the high-velocity horizon, as is the case in the calcareous areas. Therefore the material above the presumed limestone layer is likely to be both calcareous ooze and red clay and mixtures of the two; it did not necessarily lithify at the exact boundary between the two materials. Consequently the rate of deposition of the materials above the presumed limestone layer was likely to be, on the average, higher than for clay alone—more nearly in the range of 0.75 cm/1000 years.

TABLE 4.—THICKNESSES OF ORIGINAL ACCUMULATION NECESSARY TO CONSOLIDATE TO PRESENT THICKNESSES IN A CLAY-SHALE SECTION

A transition from clay to shale is assumed to take place at 300 m. Significant difference is slight if this transition is not assumed; the consolidation curve for clay shows an original cumulative thickness of 3500 m required to consolidate to a present thickness of 1500 m, whereas, with the assumption, 3600 m is required. This table (down to 2000 m) is derived from Table 1.

Depth interval in sediment (m)	Present thickness increment (m)	Original accumulation, thickness increment (m)	Cumulative thickness of original sediment (m)
0–100	100	155	155
100–200	100	190	345
200–300	100	205	550
300–400	100	220	770
400–500	100	230	1000
500–600	100	235	1235
600–700	100	240	1475
700–800	100	245	1720
800–900	100	250	1970
900–1000	100	260	2230
1000–1500	500	1370	3600
1500–2000	500	1490	5090
2000–2500	500	1600	6690
2500–3000	500	1670	8360

A refraction profile shot above this layered section would have a ray refracted along the top of the limestone layer at the speed of the lower, high-velocity layer. The next refracted ray would be in the basalt, because the shale, having a lower velocity than the overlying limestone, would not refract a ray back to the surface (Snell's Law). With the assumed velocities and thicknesses shown in Figure 10, computations indicate that the section would be thicker by 30 m than would be the case if all the material between the sediment and the basalt were of 5.5 km/sec velocity.

The above supposition of a relatively thin limestone layer overlying a lithified clay deposit is important for the geochemical balance. One bar to supposing that the second layer is a limestone is that the geochemical balance could not permit that much calcareous material

The present suppositions satisfy this demand, the results of seismic work at sea, and the probabilities of historical geology. The alternative suggestion of Revelle (1954, p. 328) that these earlier (pre-Mesozoic) deposits are not now detectable by seismic means appears to be correct if these suppositions are true.

The above hypotheses require that a number of deductions be checked from field work. One is that the underlying calcareous layer in the northeast Pacific fades out and disappears somewhere to the north and that the thickness of sediments (layers 1 and 2) is greater to the south (and into the East Pacific calcareous area), owing to the faster rate of deposition of calcareous ooze than red clay. Data available are insufficient to prove this, but the following information may be indicative:

In the Northeast Pacific red-clay area at Lat. 35° N., the total thickness of layers 1 and 2 is 700 m (Gaskell *et al.*, 1958, Station 8). South of this station, but still in the red clay area, the total average thickness is 1110 m (Raitt, 1956, Mid-Pacific Stations 1, 2, 3, 4, 6, 7; Lat. 9°–27° N.). In the Eastern Pacific equatorial area of maximum accumulation of calcareous material the total thickness of layers 1 and 2 averages 1490 m (Raitt, 1956, Mid-Pacific Station 5, Capricorn Stations 24, 25). In the calcareous area south of the equator the total average thickness is 1350 m (Raitt, 1956; Capricorn Stations 18, 19, 20, 21, 22, 23; Lat. 9°–15° S.).

Thus the little information presently available indicates a progressive average thickening of the sediment column (layers 1 and 2) to the south as expected.

Raitt has shown that the seismic velocity of volcanic material commonly found on seamounts and around volcanic islands is approximately 4.0 km/sec. The areas around most volcanic islands probably contain much volcanic detritus, if not actual flows. This material, mixed with lithified sediments, would not ordinarily be separable on the basis of velocity determinations alone. An example of this type of area is undoubtedly the archipelagic apron around the Mid-Pacific Mountains, the Cretaceous islands now sunk to depths of 1100 fathoms. Three of Raitt's stations (Mid-Pacific Stations 10, 11, and 12) were shot on this apron and show an average first-layer thickness of 0.39 km and an average second-layer thickness of 2.33 km (in which the seismic velocity was about 4.86 km/sec) for a total average thickness of 2.72 km. This total thickness is significantly greater than for the averages of the Northeast Pacific red-clay area and the East Pacific calcareous area (previously cited) and bears out the opinion of Menard (1956) that the aprons around such ancient, eroded islands should be composed of sediments, pyroclastics, and volcanic flows and should be much thicker than sediments of the deep sea.

After the publication of Raitt's (1954) seismic survey of Bikini and Sylvania, Emery *et al.* (1954, p. 131) noted (*see* Fig. 1):

> "Raitt's work, however, while clarifying certain relations has raised another problem for which at the present time there seems to be no satisfactory solution. His cross sections, showing the presence of materials of higher sound velocity (basalt) near the margins of the guyot were confirmed by dredging but in the central parts of the guyot his sections indicate from several hundred to several thousand feet of material with a sound velocity comparable to that of the unconsolidated limestones capping Bikini."

The discovery on Sylvania of partially phosphatized lower Eocene *Globigerina* ooze (Hamilton and Rex, 1959), which is almost identical to that of samples 3, 4, and 5 of this study, from the guyots of the Mid-Pacific Mountains, indicates that this material was being deposited on the surface of Sylvania during early Tertiary and furnishes a ready explanation for the velocities shown by Raitt (about 2.45 km/sec); Raitt's supposition that this material could be calcareous sand and volcanic material is thus probably entirely correct; the material probably is lithified planktonic Foraminifera and, possibly, volcanic rocks.

A third typical composite section of clay and calcareous ooze should be found in the east-central Pacific and other areas where calcareous ooze is now being deposited. If a pre-Mesozoic ocean basin occurred in the east-central Pacific (as is probable) clay must have been deposited to the Middle Cretaceous and calcareous material from then to the present. A clay-shale layer undetectable by seismic means probably underlies the present calcareous sediments.

Summary

There is a very important job to be done in computing the volume of sediments presently in the ocean basins of the world. To do this properly a much more comprehensive seismic-survey coverage of the ocean basins is needed; the surveys will have to be classified by environment into basins, ridges, continental

shelves and slopes, and similar areas that contain varying thicknesses of sediments because of their location and sedimentary history. More important, it must be decided whether or not the second layers are the lithified equivalents of the first layers and/or other, different, sediments deposited under different conditions.

A deep-sea clay should consolidate along the expectable lines for a clay, as presently understood in soil-mechanics theory. A relatively fast decrease in porosity should take place, under the buoyant weight of the mineral grains, to depths of the order of 150–700 m where, with porosities less than 45 per cent, the sediment starts to lithify into shale, and seismic results should show a marked increase in velocity and conversion to shear waves. The velocities of the elastic compressional wave should not be higher than about 4.5 km/sec in a shale section derived from a low calcium carbonate clay and will usually be much lower.

A calcareous-ooze deposit should consolidate along the lines expected of a sandy silt or a silty sand; porosity decrease should be much slower than in a clay section, so that at the same depths an unlithified calcareous section always will be more porous than a clay or silty-clay section. At about the same depths as indicated for shale (150–700 m) the ooze should lithify into a limestone because of pressure, age, and chemical factors; the depth to such a layer is probably indicated by the seismic work of Raitt (1956) in the east-central Pacific (200–600 m). The discovery that foraminiferal limestone derived solely from the lithification of calcareous ooze has velocities ranging from 1.8 km/sec to 5.5 km/sec at room temperature and pressure, plus literature values, indicates that most of the high-velocity values determined by seismic work at sea might be explained by normal deposition, consolidation, and lithification.

Geologic history indicates that the distribution of calcareous deposits during the late Cretaceous and Early Tertiary was probably more widespread than today; depositional changes from calcareous material to clay might explain some seismic sections where present red clay overlies a second layer of high velocity. The view has been expressed that the second-layer material cannot be limestone because the geochemical balance cannot tolerate that thickness and volume of calcareous material. An examination of possible seismic models in an ancient basin (*e.g.*, Fig. 10) indicates that the high velocities of a second layer are from the top of such a layer, and there is no evidence that the reported thickness of this layer is a[l]
of the same composition. On the contrar[y]
probable depositional history indicates that
lower, undetectable layer of clay shale is likel[y]
to be present in many areas. Thus, it is n[o]
longer tenable to suppose that the second
layer velocity cannot indicate limestone be
cause of the geochemical balance.

The concepts of this paper will not apply t[o]
all sections measured seismically in the oceans
As more and more evidence is derived about th[e]
sea floor, generalizations are likely to be un-
tenable. Near volcanic islands and ancien[t]
drowned islands, the sediments of the archi
pelagic aprons are likely to be thick and com
posed of pyroclastics, erosional debris, lava
flows, and clay and organic material settling
from the upper waters. The seismic picture i[s]
undoubtedly complicated in many areas (in-
cluding the deep sea) by volcanic effusions
building up into seamounts, seaknolls, or in
truded into the sediment column; fracture
zones and areas of slumping and lack of ac
cumulation are further complicating problems
Each area must be studied as an entity. Th[e]
thickness and composition of the sediments o[n]
the deep-sea floor will never be exactly know[n]
until holes are drilled in selected areas; unti
that time we can only use the best indirec
evidence available and consider all alternative

The thickness of sediments in an ancien
deep-sea basin is probably the total thickne
of the upper unconsolidated layer and the lowe
higher velocity layer; in short: the total thick
ness of layers 1 and 2 of seismic literature.
measurement of sediment thickness at an
time horizon of the past, or future, should
show about the same "anomalously thin" valu
for the topmost part of the slowly consolidatin
sediment. The presence of intermediate velocity
layers between the "unconsolidated" sedimen
and the basaltic basement rock can thus be e
plained in many areas as normal, expecte
consolidation and lithification of the prese
types of sediment in an ancient ocean basin.
this is true then the geochemical balance callin
for large amounts of solids on the sea floo
recent seismic discoveries in the ocean basin
and the consolidation theories of soil-mechani
engineering fall approximately into line.

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Appendix A. Hydrostatic Uplift

The question of hydrostatic uplift (buoyancy of mineral grains due to Archimedes' principle) is important in computing pressures within the sediment. The submerged weight of the grains (the intergranular pressure) is less if uplift is figured; this pressure is independent of the depth of the water as long as free movement of the water can take place so that the excess pore-water pressure can become zero. The rate of pressure changes in the pore water depends primarily on the permeability of the sediment or rock and resistance of the grains to deformation and repacking (Emery and Rittenberg, 1952, p. 761). All sediment is permeable, and even unfractured rock cannot be regarded as impervious (Peck *et al.*, 1953, p. 58). This means that over long periods of time the pressure adjustment is made to the hydrostatic case (excess pore-water pressure equal to zero), which is the situation in most oil-field reservoirs (Levorsen, 1954, p. 386).

In 1936, Terzaghi studied hydrostatic uplift in plastic clays and concrete and came to several important conclusions: (1) test results disclosed that an increase in pore-water pressure (above the normal hydrostatic pressure) produced no measurable compression of a clay or of concrete regardless of the compressibility of the material, (2) tests on a stiff, plastic Tertiary clay showed that hydrostatic uplift was 95–99 per cent effective despite low permeability of the clay, and (3) in concrete with porosities as low as 13 per cent, hydrostatic uplift was practically 100 per cent.

In the field of soil mechanics 100 per cent hydrostatic uplift is allowed in foundation studies. The question then follows: at what point (if any) in the consolidation and consequent lithification of a sediment does hydrostatic uplift cease to be effective, and the overburden pressure (effective in causing deformation) equal the sum of the intergranular pressure and the pressure due to the weight of the water?

Jones (1944, p. 142) noted that hydrostatic uplift is effective for high void ratios but doubted that it is effective for low ratios (on the order of 0.1); he believed that when pressure is calculated hydrostatic uplift should be figured for the high porosities and none for the low, with some intermediate figure for the moderate ratios. Skempton (1953, p. 52) thought that hydrostatic uplift is effective down to at least 3000 feet depth or to about 100 kg/cm² of intergranular pressure. Hedberg (1936), in computing overburden pressures for Venezuelan shales, considered that no effective hydrostatic uplift occurred.

Skeels (*in* LeRoy, 1950) amassed a great deal of data on the variations of density and porosity with depth in shale sections; the relationship of porosity with depth in shale is thus well known. This porosity reduction could be a consequence of the submerged weight of the mineral grains (allowing for hydrostatic uplift) or could be due to the weight of the minerals and the water. The writer made an *e-log p* plot for both contingencies and computed the depth-porosity-pressure relationships. Figure 3 illustrates the pressure-porosity relationships for Skeels' data with and without hydrostatic uplift and for red clay and *Globigerina* ooze with hydrostatic uplift; Figure 4 shows the depth-porosity relationships for Skeels' shale, red clay, and *Globigerina* ooze with hydrostatic uplift. In oil-field reservoirs pressures are usually hydrostatic and not the combined pressure induced by the weight of the rock and the water. In some sealed reservoirs the pressure is that of the rock and the water (Levorsen, 1954). This indicates that in normal situations the pressure induced at any level in a rock section does not include the weight of the overlying water as effective pressure on the mineral grains which would cause consolidation. Somewhere near the bottom of a thick shale section, where the porosity is very low, hydrostatic uplift may not be effective, but the writer thinks that the higher porosity sections with which this study is concerned should be considered to have effective hydrostatic uplift of the mineral grains. For the purposes of this paper all pressures are computed with full hydrostatic uplift unless otherwise stated; this is the minimum case in explaining high seismic velocities because velocities are likely to be higher with higher pressures.

Many investigators have noted that porosity does not vary appreciably in the uppermost layers of many sediments in the sea (Shepard and Moore, 1955, p. 1580; Arrhenius, 1952) or in clay layers under foundations. Terzaghi (*in* Skempton, 1953) noted clays that could support loads up to about 0.1 kg/cm² before the sediment structure collapsed and deformation began. Terzaghi considered that this phenomenon is due to the intermolecular bonding in clays in areas of slow deposition resulting in a structure that could support small loads; therefore little consolidation and reduction occurred in porosity to shallow depths.

The porosity-depth curve for Tertiary sediments and shale of the Gulf Coast (Dickinson, 1953) is close to the computed curve for clay (grain density of 2.70 gr/cc) with 100 per cent hydrostatic uplift (Fig. 4). These Gulf Coast sediments and rocks were deposited in continuous sequence without major unconformities and usually have a "normal" (hydrostatic) reservoir pressure. The position of the curve (Fig. 4) and the reservoir pressures argue for full hydrostatic uplift of the mineral particles of the rock. Curves for shale should lie on the low-porosity side of those for clay: cementation should effect a reduction of porosity not present in clays.

Appendix B. Construction of Tables 1, 2, and 4

In constructing Tables 1 and 2, typical properties of deep-sea clay and calcareous ooze were used as starting values for the sediment at the surface of the present-day sea floor. Using standard soil-mechanics methods the tables were then constructed in the following manner:

(1) Using the submerged unit weight of the surface sediment (the *in situ* saturated density minus the density of the interstitial water) the weight of a column of sediment 10 m high was computed and used for the first increment of interval pressure; this assumes no deformation in the top 10 m of deep-sea sediment, as is approximately true (Arrhenius, 1952, p. 90).

(2) Using the cumulative pressure (which is the same as the interval pressure for the first increment only) the *e-log p* curve of Figure 2 (for colloidal clay or *Globigerina* ooze) was entered, and the void ratio *e* picked off for this pressure.

(3) The porosity, *n*, was computed using the relationship: $n = e/(1 + e)$

(4) The *in situ* density of the sediment at the new void ratio and porosity was computed using the relationship:

$$D_{sed} = n(D_w) + (1 - n)D_{sol},$$

where D_{sed} equals the *in situ* density of the sediment, D_w equals the density of the sea water at about 5 km depth, and D_{sol} equals the average density of the mineral grains of the sediment.

(5) For successively deeper increments an extrapolated submerged unit weight for the next lower increment was used to find the interval pressure; the cumulative pressure is the additive weights of the interval weights.

(6) In constructing Table 1, which illustrates the consolidation of a clay-shale section, the writer used the *e-log p* curve for colloidal clay (Fig. 2) from 0.1 to 19.5 kg/cm². At the latter pressure the porosity is 41 per cent, and the depth below the water-sediment interface is 300 m. This is about the depth at which theory and actual upper porosities of shale sections indicate a beginning of lithification into shale. From 300 m to greater depths the *e-log p* curve for shale with hydrostatic uplift (computed from data of Skeels) was used in computing Table 1. Part of the shale curve is illustrated in Figure 5. The appearance of the curves of Figure 4 suggests that the above assumptions are reasonable.

In computing Table 2 for *Globigerina* ooze the *e-log p* data from Laughton (1957) was used to illustrate the consolidation of this typical deep-sea sediment.

Clays and oozes will vary from the computed consolidations depending on their individual characteristics, but soil-mechanics experience, especially with clays (*see* Fig. 2), indicates that deep-sea sediments will vary only slightly from these determinations in the upper portion of the geologic

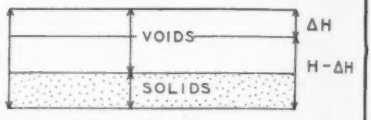


Figure 11.—Diagram Illustrating Settlement Computations

column; this variance should not be sufficient to invalidate the general conclusions.

From the tables one can construct curves showing the variations of pressure, density, and porosity with each other and with depth in the sediments (Figs. 3, 4). The writer does not think that these tables or figures are always valid for the deeper parts of the sediment column; however, they were computed to a depth of 2 km and drawn so that the data could be compared to similar curves from other papers.

If the thickness of a clay layer, *H*, is known, as well as the void ratio before and after consolidation, it is possible to compute the amount of settlement to be expected (*e.g.*, Tschebotarioff, 1951, p. 105, Fig. 11).

Settlement is computed by the equation:

$$\Delta H = H \frac{e_o - e_f}{1 + e_o},$$

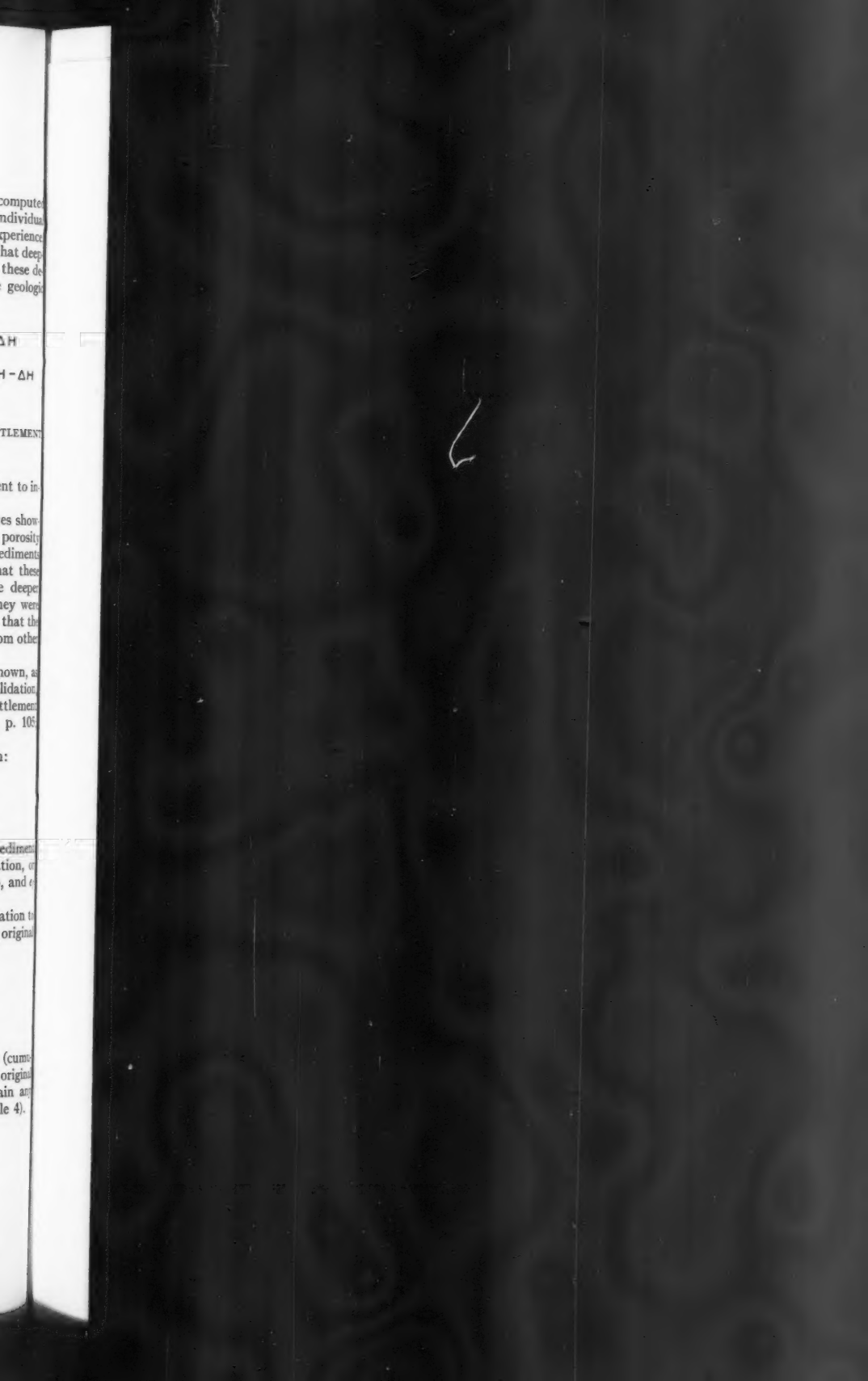
where *H* equals the original thickness of a sediment layer, ΔH equals the amount of consolidation, or settlement, e_o equals the original void ratio, and e_f equals the final void ratio.

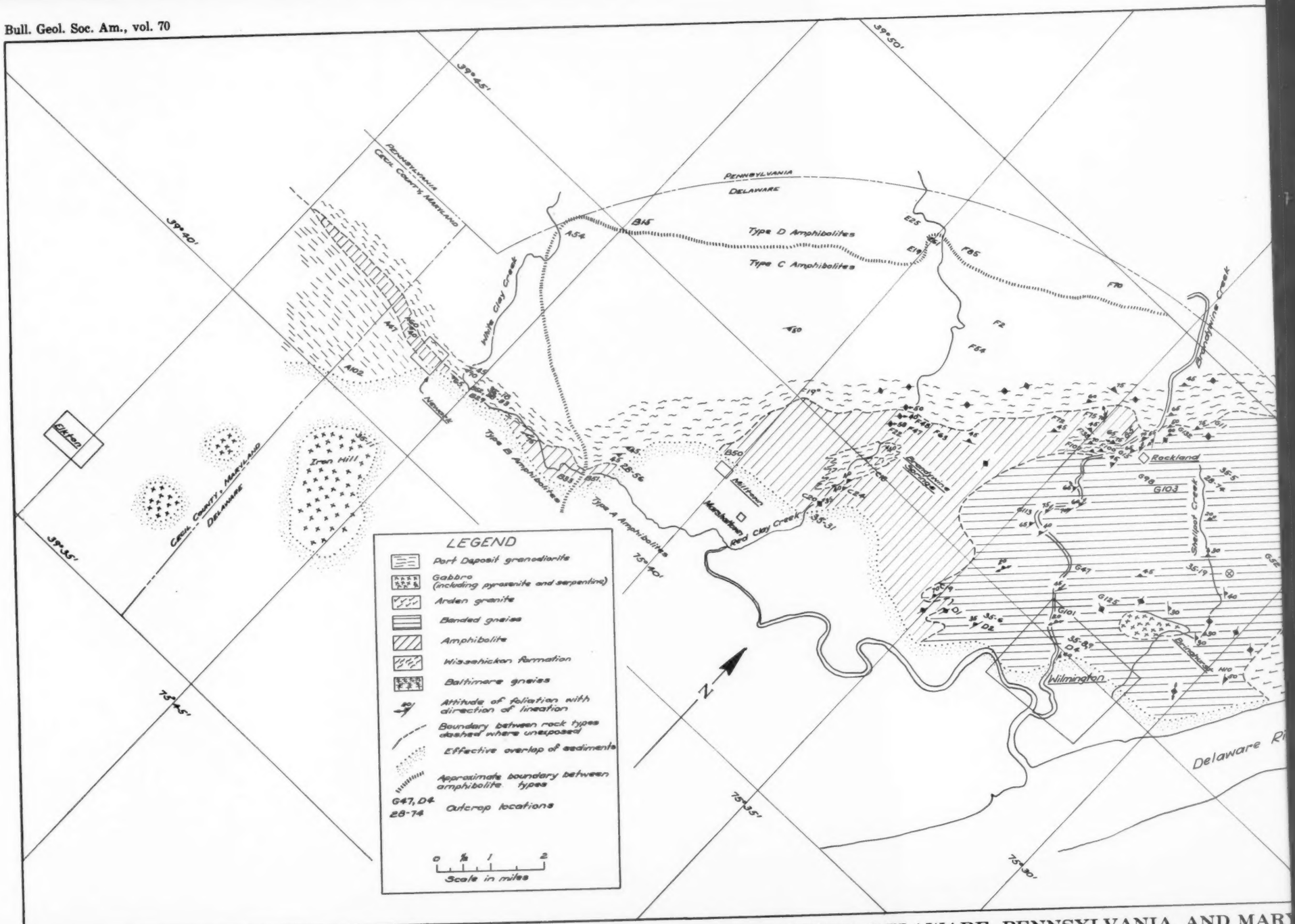
It is convenient for the present investigation to substitute into the equation to find *H*, the original thickness,

$$H = \left(\frac{1 + e_o}{1 + e_f}\right)(H - \Delta H),$$

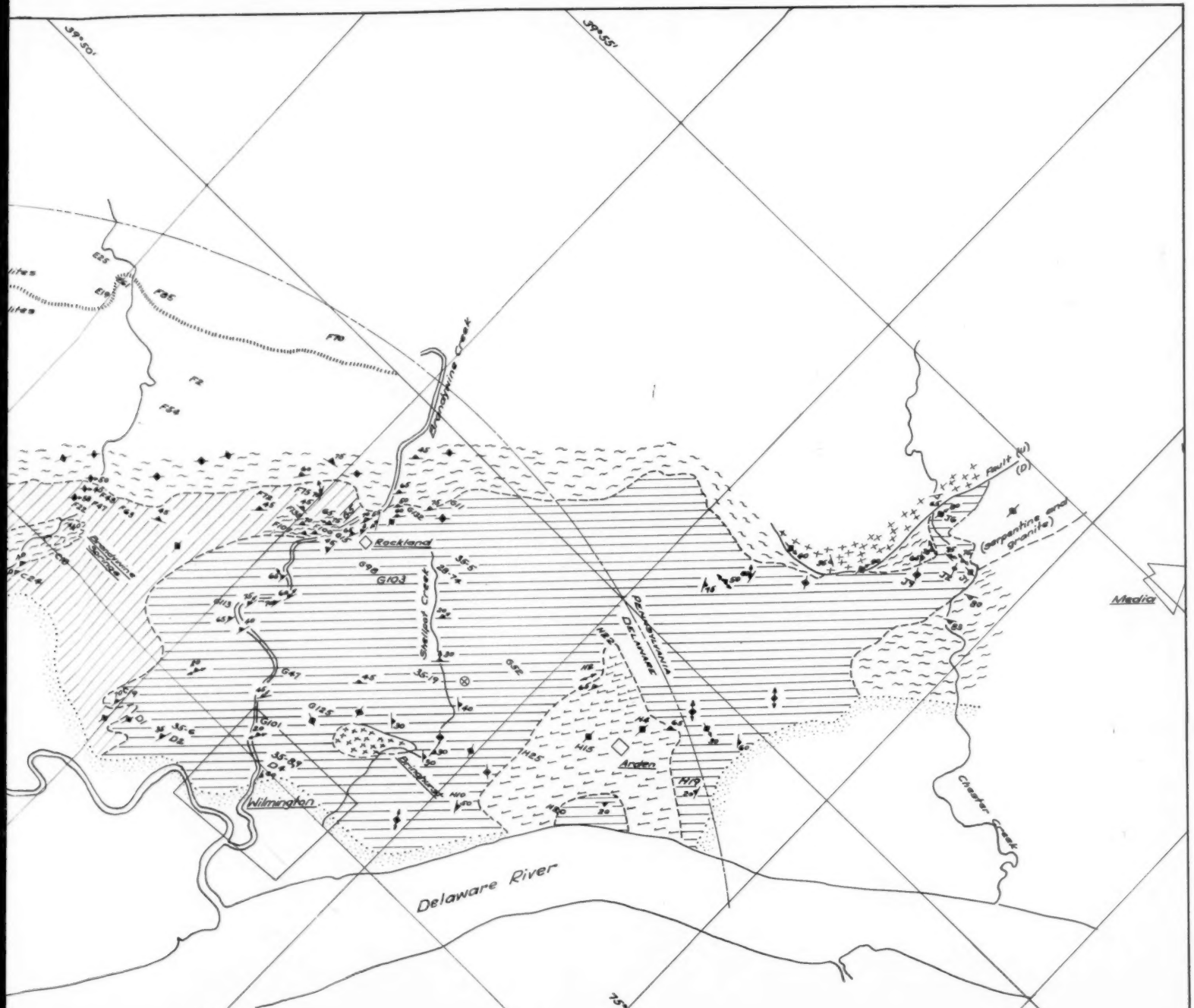
where $H - \Delta H$ equals present thickness.

The addition of the expanded increments (cumulative *H*) shows approximately how much original sediment deposition was necessary to obtain any stated present thickness of sediments (Table 4).





MAP OF WILMINGTON COMPLEX, DELAWARE, PENNSYLVANIA, AND MARY



X, DELAWARE, PENNSYLVANIA, AND MARYLAND

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PETROLOGY AND METAMORPHISM OF THE WILMINGTON COMPLEX, DELAWARE, PENNSYLVANIA, AND MARYLAND

BY RICHARD F. WARD

ABSTRACT

The Wilmington complex occupies about 100 square miles in Delaware and adjacent Pennsylvania and Maryland and is bordered by the Wissahickon formation (lower Paleozoic?) and the Port Deposit granodiorite. Its southern edge is covered by Coastal Plain sediments. It consists of a banded gneiss comprising mafic bands of calcic plagioclase, hypersthene, augite, and commonly hornblende and thick felsic bands of quartz and andesine with minor pyroxene. In the central portion the gneiss has been invaded by a granite, probably metasomatic, consisting of orthoclase, oligoclase-andesine, biotite, and pyroxene. To the southwest the banded gneiss grades into amphibolites. Amphibolites near the gneiss contain calcic plagioclase, hornblende, and traces of hypersthene. Those farther away are hypersthene-free, and the plagioclase is replaced in part by epidote. Small bodies of amphibolite within the adjacent schist show similar characteristics.

Veins of granite, apparently related to the Port Deposit granodiorite, intrude the southwestern part of the complex. A coarse-grained undeformed gabbro stock has intruded the gneiss north of Wilmington; a similar body of gabbro occurs as a group of low hills surrounded by Coastal Plain sediments.

The relationship between the chemical composition of the rocks and the mineral phases present indicates that the central granite body and the banded gneiss crystallized or recrystallized under granulite-facies conditions. The amphibolite near the gneiss represents upper almandine amphibolite facies (probably almandine-sillimanite subfacies), and the more distant amphibolites, having undergone greater stress, probably represent the almandine-diopside-hornblende subfacies. The complex is probably along the axis of most intense metamorphism of the Appalachian(?) Revolution.

Structurally, the complex is characterized by the banding (S_1) and foliation (S_2) which are commonly, but not always, parallel. Structures in the complex generally parallel structures in the Wissahickon formation, indicating simultaneous deformation.

Recrystallization of the mafic portion of the banded gneiss was accompanied by the introduction, probably metasomatic, of the felsic material. Mafic bands may be transformed into felsic bands by the introduction of Si and Na and the removal of Ca, Fe, Mg, and Al representing a transfer of approximately 7 per cent material by volume. This "potassium-free granitization" may have a definite spatial relationship to the true granitization that produced the central granite. The origin of the mafic materials which comprise the gneiss and the amphibolites may have been (1) sedimentary (impure limestone); (2) volcanic (basaltic tuffs or flows); (3) hypabyssal (a lopolith); or (4) plutonic (a gabbro). Volcanic origin is tentatively suggested. If this is the case the rocks may be correlative with metabasalts of nearby Cecil County, Maryland, and with the Catoctin-South Mountain metabasalts in western Maryland, both of which underlie the Glenarm series.

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INTRODUCTION

Geography

The area of study is in the northern portion of the Appalachian Piedmont province, roughly halfway between Philadelphia and Baltimore. The rocks are disposed parallel to the regional strike from near Media, Pennsylvania, through northern Delaware to Elkton, Maryland (Fig 1).

The outcrop area, approximately 100 square miles, ranges in altitude from less than 100 feet near the Fall Line to more than 400 feet along the divides. Exposures are good along major streams and their swift-running tributaries. The divide areas are level, and outcrops are scarce. However, float and boulders are abundant, and the underlying lithology may be inferred even though the structure is hidden. There are many quarry exposures, for the mafic rocks make excellent crushed stone.

In the Fall Zone the sedimentary cover is irregular, and bedrock is commonly exposed in stream beds for some distance downstream from the uppermost sediments. For this reason the maps presented herein show the effective overlap of the sediments, that is, the line below which no reliable exposures of crystalline rocks can be found.

Geologic Relationships and Previous Investigations

A belt of igneous and igneous-appearing rocks lies southeast of the Glenarm series between Philadelphia and the Susquehanna River. Postel (1940) and Weiss (1947) have studied in detail the northeastern end of the belt. Hershey (1937) has described the Port Deposit granodiorite at the southern end of the area as a composite batholith intrusive into the Glenarm series, and Marshall (1937) has described the metavolcanic rocks of Cecil County, Maryland, as a roof pendant in the batholith. In the central part of the belt lies one of the largest areas of "post-Glenarm gabbro" of the existing geologic maps (Bascom, 1902; Bascom and others, 1909; Bascom and Miller, 1920; Bascom and Stose, 1932). This body is the subject of the present investigation.

Chester (1890) described in detail the textures and minerals of the "gabbros of Delaware." Bascom and others (1909), Bascom and Miller (1920), and Bascom and Stose (1932) mapped the body to be discussed herein as gabbro, undifferentiated. The texts accompanying the maps, however, refer to a wide range of composition and texture. As part of an investigation of the composition and optical properties of the

major rock-forming minerals of this area being one at Bryn Mawr College, specimens of the orthopyroxenes (Clavan, McNabb, and Watson, 1954), clinopyroxenes (Norton and sists of a dark-colored bluish rock which comprises alternating lighter and darker bands. The felsic material makes up about 75 per cent of the volume of the gneiss. The darker, mafic

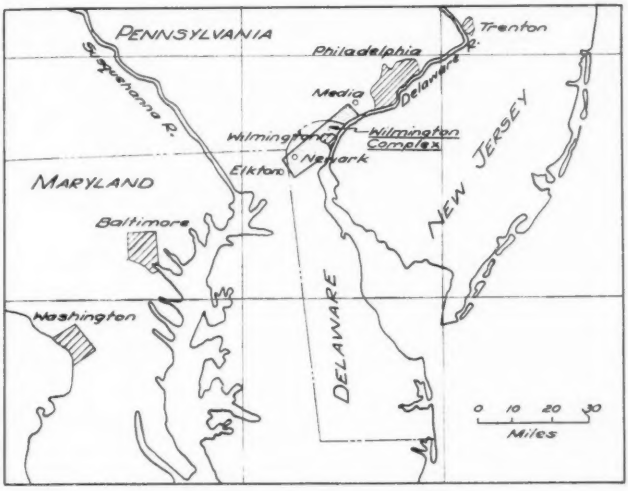

FIGURE 1.—LOCATION MAP

Clavan, 1958), and hornblendes (Rosenzweig and Watson, 1954) from these rocks have been described. Ward and Groot (1957) have described several specimens from the complex and from the adjacent Wissahickon formation in a recent evaluation of potential quarry sites in Delaware.

The degree of metamorphism in the Wissahickon formation immediately north and northeast of this complex increases as the complex is approached. The adjacent schists show sillimanite grade (McKinstry, 1949) or almandine-sillimanite subfacies of the amphibolite facies (Wyckoff, 1952).

Present Investigation

The present investigation was undertaken to determine the nature and relationships of the rock types which make up the complex, to determine the age relationships with the surrounding metasediments, and to determine the physical conditions attendant upon its formation.

The complex comprises banded gneiss, Arden granite, amphibolite, gabbro, and the Port Deposit granodiorite (Fig. 2):

BANDED GNEISS: Most of the complex consists of a dark-colored bluish rock which comprises alternating lighter and darker bands. The felsic material makes up about 75 per cent of the volume of the gneiss. The darker, mafic bands range in thickness from a few inches to several feet and contain approximately 50 per cent plagioclase (An_{33} to An_{64}) and subequal portions of hypersthene and augite. Replacement of pyroxene by hornblende is widespread. The felsic bands comprise subequal amounts of quartz and plagioclase (An_{30} to An_{50}) along with about 10 per cent pyroxene. Hornblende and biotite, commonly in small quantities, are widely distributed.

ARDEN GRANITE: In the central portion the banded gneiss grades into, or is intruded by, a rock which ranges from a composition near that of the felsic bands of the gneiss (andesine-pyroxene-quartz) to granite (orthoclase-oligoclase-quartz-biotite).

AMPHIBOLITE: To the south and west the banded gneiss grades into hornblende-plagioclase rocks which lack the distinctly banded character of the gneiss but show considerable variation in the proportions of the major constituents in adjacent layers. Near the western end of the complex specimens occur which contain andesine-epidote instead of more calcic plagioclase and augite-hornblende instead of hornblende. Amphibolites within the adjacent Wissahickon formation are described and admit

ot the same twofold division as the amphibolites of the complex.

GABBRO: A small stock of coarse massive gabbro intrudes the banded gneiss at Bring-

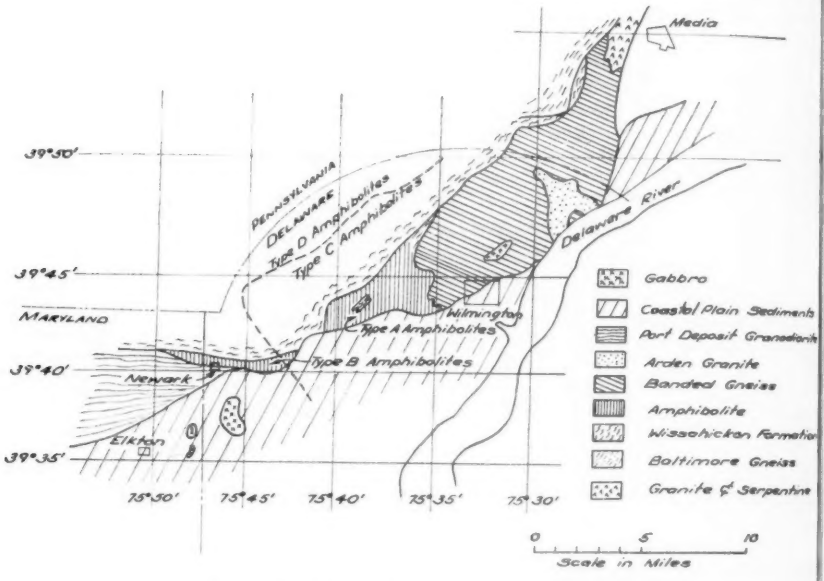

FIGURE 2.—SKETCH MAP OF WILMINGTON COMPLEX

hurst, just north of Wilmington, Delaware. Several xenoliths of mafic bands from the gneiss are enclosed in this gabbro near the contact.

Between Newark, Delaware, and Elkton, Maryland, are exposed four separate areas of coarse-grained rock ranging from norite through pyroxenite to peridotite and serpentine. The feldspar-bearing portions are largely uralitized. These bodies, called the Iron Hill stocks, are completely surrounded by Coastal Plain sediments; consequently their geologic relationships are poorly known.

PORT DEPOSIT GRANODIORITE: The Port Deposit granodiorite is a microcline-bearing rock which occupies a large area southwest of the Wilmington complex. Locally it intrudes the amphibolites of the complex; it is treated in this report only in that connection.

Methods of Investigation

The field mapping for this report was done during parts of 1954, 1955, 1956, and 1957.

Plagioclase determinations were made with index oils in white light on 010 cleavage flakes, employing the curves of Tsuboi (1923). Wh possible, these were checked against values o tained by extinction angles in individuals sho ing albite or Carlsbad-albite twinning. Accura of the determinations is approximately ± A

Important specimens were analyzed micr metrically with the equipment and metho outlined by Chayes (1949); 1000 to 2000 poin were counted, depending upon grain size an quality of the thin section. The compositions o other slides were estimated.

Six of the chemical analyses given were do at the University of Minnesota Rock Analys Laboratory. The seventh was done at th Chemistry Department of Bryn Mawr Colleg

The locations of all outcrops and thin section mentioned in the text are shown on Plate 1. Fo every outcrop mentioned by number in th text, one or more modes are given in the ap propriate table (except G113, from which spec mens could not be obtained). In cases whe more than one thin section was prepared fro an outcrop these are distinguished by lower-ca letters, such as A25a, A25b, and A25c.

Acknowledgments

I should like to acknowledge my deep i debtedness to Dr. Dorothy Wyckoff for her a

vice on all phases of the work. Dr. Edward H. Watson critically reviewed many outcrops and field data. Dr. H. E. McKinstry, Dr. Anna Hietanen Makela, and Dr. H. R. Gault read the manuscript and pointed out its shortcomings. I alone am responsible for any remaining defects.

The Delaware Geological Survey supported the project, including six chemical analyses and more than 200 thin sections. The field and laboratory work were done as the pressure of my other duties at the Survey permitted. Publication of the results has been authorized by the State Geologist of Delaware, Dr. J. J. Groot.

Mineralogy

Twenty-five mineral species have been recognized in these rocks. Quartz, plagioclase, hornblende, hypersthene, augite, biotite, and potassium feldspar are the most significant minerals present and account for more than 95 per cent of all of the rocks. Epidote, chlorite, magnetite, scapolite, garnet, zircon, apatite, monazite, kyanite, allanite, sphene, rutile, calcite, muscovite, serpentine, chromite, spinel, and tourmaline occur in small amounts.

Detailed physical, optical, and chemical data for three critical minerals from portions of the complex were available to the author from five samples of clinopyroxene (Norton and Clavan, 1958), five samples of orthopyroxene from the same specimens (Clavan, McNabb, and Watson, 1954), and four samples of hornblende (Rosenzweig and Watson, 1954). One hornblende is from a specimen used by Norton and Clavan.

QUARTZ: Most of the quartz, which occurs in equant or slightly elongated grains, exhibits undulatory extinction. Those grains showing most strain are peripherally granulated. In the banded gneiss and the Arden granite the quartz appears bluish in reflected light. Such color is common in metagabbros and charnockites which occur in terrains of high-grade regional metamorphism (Turner, 1948; Johannsen, 1932, p. 133–137).

The specimens of blue quartz contain a few very small acicular inclusions with high refractive index—possibly rutile. These show no regular crystallographic orientation. In addition they contain very small, irregular inclusions of an isotropic material with an index of refraction much lower than quartz, possibly opal produced by the action of trapped liquid in the quartz. However, neither of these materials seems to be sufficiently abundant to account for the even bluish color seen in reflected light.

POTASSIUM FELDSPAR: Potassium feldspar occurs in the Port Deposit granodiorite and associated veins as microcline which is pink or greenish in hand specimen and has no unusual microscopic characteristics.

In the few specimens of amphibolite which contain potassium feldspar, it is interstitial orthoclase, with a large optic angle. No twinning has been detected.

The Arden granite contains considerable potassium feldspar, which occurs in the following ways:

(1) As antiperthite in sodic plagioclase. No twinning has been detected in this potassium feldspar, and it is presumed to be orthoclase. Individual grains are too small to permit detailed optical tests.

(2) As perthitic individuals of orthoclase. Carlsbad twinning is common, particularly in the large grains (1 cm or larger). The optic angle is large, and the extinction angle does not exceed 6°.

(3) As individuals present in a few specimens, which are very similar to those of the second group, except that the optic angle is small to very small. Measurement with the universal stage (in white light) has revealed grains with (−) 2V of 8°, 41° and one in which no separation of optic axes could be detected. Within each of these grains, however, some variation is present which appears irregular rather than zonal.

PLAGIOCLASE: Plagioclase ranges from An_{15} to An_{90}. Zoning is either absent or obscured by undulatory extinction in most of the rock types. An exception occurs in amphibolites within the Wissahickon formation in which epidote accompanies the plagioclase. In these, nearly every grain shows a core of more calcic composition than the outer portion.

Plagioclase twinning in the banded gneiss, the Arden granite, and the amphibolites is simple and poorly defined or absent. This is considered typical of recrystallized plagioclase (Turner, 1951; Gorai, 1951). By contrast, in the gabbroic rocks of the Iron Hill and Bringhurst stocks and in the xenoliths of gneiss within the Bringhurst gabbro, twinning is universal, well defined, and in most cases complex. Examination with a 4-axis universal stage, using the Rittman zone method (Emmons, 1943) and the Nikitin-Fedorow method (Reinhard, 1931), has revealed a preponderance of albite-Carls-

bad, albite-Ala and Manebach-Ala twins in plagioclase from Iron Hill.

The plagioclase is fresh in all unweathered specimens except where the Port Deposit granitive and moderate, 50°–60° (estimated). Measured values (Clavan, McNabb, and Watson, 1954, p. 576) range from 51° to 64°, with an average of 56°.

TABLE 1.—CHEMICAL ANALYSES OF HYPERSTHENE

(From Clavan, McNabb, and Watson, 1954; Walter Clavan, analyst)

	35-5*	35-6	35-8	35-9	35-19
SiO_2	50.70	50.60	51.43	52.00	52.80
Al_2O_3	0.05	0.04	0.06	0.57	0.00
Fe_2O_3	1.18	1.44	1.39	1.34	1.67
FeO	25.65	26.02	23.60	22.46	18.65
MgO	20.29	19.49	20.69	21.62	24.95
CaO	0.66	0.87	1.02	0.75	1.05
Na_2O	0.06	0.07	0.46	0.46	0.44
K_2O	0.05	0.06	0.06	0.07	0.01
H_2O	0.22	0.28	0.32	0.28	0.14
TiO_2	0.17	0.17	0.18	0.14	0.09
Cr_2O_3	0.00	0.004	0.00	0.003	0.02
NiO	0.00	0.00	0.00	0.00	0.01
MnO	0.75	0.95	0.73	0.57	0.43
Total	99.78	99.99	99.94	100.26	100.26
2V	−52°	−51°	−55°	−57°	−64°
alpha	1.702	1.703	1.699	1.695	1.685
beta	1.711	1.713	1.708	1.705	1.695
gamma	1.716	1.718	1.713	1.710	1.700
$\gamma - \alpha$	0.014	0.015	0.014	0.015	0.015

* Specimen locations shown on Plate 1

35-5 From mafic band in banded gneiss 4 miles north of Wilmington, Delaware

35-6 From felsic band in banded gneiss exposed in park at Franklin and Sycamore streets, Wilmington

35-8 From felsic band in banded gneiss at abandoned quarry on east bank of Brandywine Creek in Wilmington

35-9 From mafic band adjacent to felsic band from which specimen 35-8 was taken

35-19 From boulder of mafic band in banded gneiss 3 miles north of Wilmington

odiorite invades the amphibolite. In these instances the plagioclase of the granodiorite and the adjacent amphibolite appears to have been altered to clay minerals by hydrothermal solutions.

HYPERSTHENE: Five analyzed samples of hypersthene (Clavan, McNabb, and Watson, 1954) are available from the gneiss. Three are from mafic bands and two from felsic bands.

In thin section the hypersthene is very weakly pleochroic to nonpleochroic, which makes it difficult to distinguish from associated augite. Thick fragments, however, have marked pleochroism: X is salmon pink, Y is yellow brown, and Z is light green.

In all hypersthene examined 2V was negative and moderate, 50°–60° (estimated). Measured values (Clavan, McNabb, and Watson, 1954, p. 576) range from 51° to 64°, with an average of 56°.

Chemical compositions of these specimens (Table 1) range from $En_{55}Fs_{43}Wo_2$ to $En_{69}Fs_{29}Wo_2$.

AUGITE: Norton and Clavan (1958) give a detailed account of the chemistry, optical properties, and internal structure of the clinopyroxenes of the complex and associated rocks. The five analyzed samples from the banded gneiss are from the same specimens used in the study of hypersthene (Clavan, McNabb, and Watson, 1954).

In thin section the augite is colorless and nonpleochroic or faintly pleochroic colorless to pale green. Even in thicker fragments, the pleochroism is not strong enough to permit determination of the pleochroic formula.

TABLE 2.—CHEMICAL ANALYSES OF AUGITE

(From Norton and Clavan, 1958; Walter Clavan, analyst) *See* footnote to Table 1 for identification of specimen numbers.

	35-5	35-6	35-8	35-9	35-19
SiO_2	51.03	51.00	50.17	51.55	51.43
Al_2O_3	2.40	2.49	2.38	2.44	2.63
Fe_2O_3	1.88	2.03	3.39	2.03	1.70
FeO	8.97	9.31	8.05	7.28	5.92
MgO	13.11	12.53	13.22	13.65	14.83
CaO	21.75	21.08	21.10	21.73	22.18
Na_2O	0.43	0.47	0.49	0.53	0.44
K_2O	0.00	0.00	0.00	0.00	0.00
H_2O	0.24	0.37	0.18	0.26	0.31
TiO_2	0.28	0.27	0.28	0.31	0.28
Cr_2O_3	0.00	0.00	0.00	0.01	0.04
NiO	0.01	0.01	0.01	0.01	0.01
MnO	0.35	0.45	0.34	0.24	0.19
Total	100.45	100.01	99.61	100.03	99.96
(+)2V	47°	44°	49°	47°	48°
Z to *c*	43°	40°	42°	40°	39°
alpha	1.687	1.692	1.687	1.691	1.686
beta	1.691	1.696	1.691	1.695	1.691
gamma	1.713	1.721	1.712	1.717	1.714
$\gamma - \alpha$	0.026	0.029	0.025	0.026	0.028

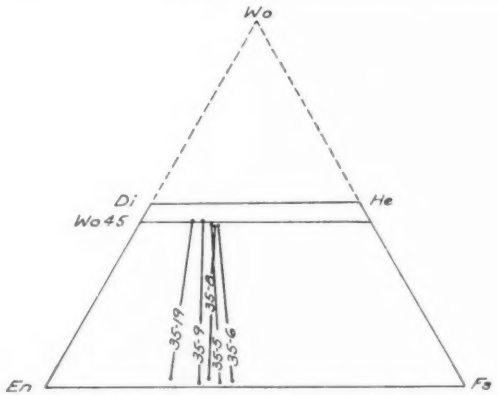

FIGURE 3.—TRIANGULAR PLOT OF ANALYZED PYROXENES

Wo, wollastonite; En, enstatite; Fs, ferrosilite (after Hess, 1941); Di, diopside; He, hedenbergite

The optic angle, 2V, is positive and moderate, 40° to 55° (estimated). Values given by Norton and Claven (1958) range from 44° to 48°, with an average of 45°.

Chemical compositions of the five analyzed specimens (Table 2) range from $Wo_{46}En_{42}Fs_{12}$ to $Wo_{44}En_{38}Fs_{19}$ and fall in or at the edge of the field of clinopyroxenes for which Hess (1941, Fig. 1) has proposed the name salite to designate clinopyroxenes intermediate in composition between diopside and hedenbergite. Hess notes (1941, p. 518):

"The majority of natural occurrences of diopside-hedenbergite pyroxenes are from contact meta-

TABLE 3.—CLASSIFICATION OF HORNBLENDES

Type	Pleochroism	Occurrence	Comment
Brown-green hornblende	X, pale yellow Y, yellow green Z, brown green	Banded gneiss	For chemical, optical data *see* 35-5, Table 4
Dark-green hornblende	X, yellow green Y, light green Z, dark green	Amphibolites	For chemical, optical data *see* 35-31, Table 4
Blue-green hornblende	X, pale yellow Y, light green Z, blue green	"Granitized" rocks	For chemical, optical data *see* 35-10, Table 4
Light-blue-green hornblende	X, pale yellow Y, light green Z, light blue green	Uralite gabbro	For chemical, optical data *see* 35-11, Table 4
Colorless	Nonpleochroic	Uralite gabbro	No chemical data

TABLE 4.—CHEMICAL ANALYSES OF HORNBLENDE

(From Rosenzweig and Watson, 1954; A. Rosenzweig, analyst)

	35-5*	35-10	35-11	35-31
SiO_2	44.94	45.09	48.71	44.73
Al_2O_3	11.24	10.43	9.48	11.54
TiO_2	1.79	0.84	0.32	1.32
Fe_2O_3	2.87	3.90	2.33	2.87
FeO	12.61	10.14	9.12	11.07
MgO	11.98	12.52	14.43	11.89
MnO	0.22	0.24	0.23	0.40
CaO	11.61	12.29	11.93	11.83
Na_2O	1.86	1.60	1.16	1.45
K_2O	0.47	0.78	0.15	0.60
H_2O	1.26	1.76	1.83	1.79
F	0.24	0.28	0.23	0.16
Total	101.09	99.87	99.92	99.65
2V	−72°	−72°	−78°	−78°
Z to c	16°	17°	18°	19°
alpha	1.653	1.650	1.638	1.654
beta	1.668	1.662	1.652	1.666
gamma	1.677	1.669	1.661	1.674
$\gamma - \alpha$	0.024	0.019	0.023	0.020

* Specimen locations shown on Plate 1

35-5 From mafic band in banded gneiss 4 miles north of Wilmington, Delaware

35-10 From group *B* amphibolite exposed in White Clay Creek 1 mile east of Newark, Delaware

35-11 From uralite gabbro of Iron Hill stock exposed in Christiana Creek 2 miles south of Newa Delaware

35-31 From group *A* amphibolite exposed in Baltimore and Ohio Railroad cut half a mile northeast Marshaltown, Delaware

morphic rocks and from veins. These crystallize at considerably lower temperature than most igneous clinopyroxenes . . . . The boundary, therefore, between salite and augite is placed at Wo_{45} because most augites from igneous rocks contain less calcium silicate than Wo_{45} and most contact metamorphic and vein salites more calcium silicate than Wo_{45}."

The compositions and Hess' boundaries are plotted in Figure 3. The relationship between the augite and the hypersthene is indicated by lines joining the pyroxenes from the same specimen.

The clinopyroxenes of the complex are almost certainly not a product of contact metamorphism but are rather a result of high-grade regional metamorphism. It is nevertheless significant that Hess does not believe that pyroxenes of this composition are likely to result from ordinary magmatism. As with the orthopyroxenes, there appears to be no significant difference in composition between those of the mafic bands and the felsic bands in the banded gneiss.

The name augite is used throughout the present paper, not out of disagreement with the system proposed by Hess, but rather because salite is an unfamiliar term and the terms salitic augite or diopsidic augite are cumbersome.

HORNBLENDE: The hornblendes of the complex are classified in Table 3 on the basis of color on Z. The specimens have been compared with those of Rosenzweig and Watson (1954), whose terms have been adopted.

The blue-green hornblende is concentrated in the extreme southern portion of the complex near the Port Deposit granodiorite and along the northern contact. It has been interpreted as resulting from alkali metasomatism (Rosenzweig and Watson, 1954). The chemical analyses (Table 4) and geologic relationships appear to confirm this. Seitsaari (1953), in a discussion of blue-green hornblendes from Finland, notes that they are indicative of lower-grade metamorphic conditions than more deeply colored varieties. While both sets of conclusions appear to be applicable to this area, these data are hardly a sufficient basis for any wide generalizations.

The actinolitic light-blue-green and colorless hornblendes are restricted to the uralites of the Iron Hill stocks. The two varieties are intimately associated and gradational.

BIOTITE: Throughout the complex the biotite is the common brown variety, which is found also in the adjacent Wissahickon schist. In some specimens of the granodiorite the brown biotite is peripherally altered to green biotite, and in many specimens of the granodiorite, green biotite occurs alone.

brown biotite: X = yellow
Y = Z = reddish brown to dark brown
green biotite: X = yellow
Y = Z = olive green to black

No chemical analyses are available.

EPIDOTE: Epidote is found in some amphibolites, the Port Deposit granodiorite, and the uralitized gabbro of the Iron Hill stock. It appears monoclinic and is therefore assigned to the clinozoisite-pistacite series except in the gabbro where it is orthorhombic and probably zoisite.

MAGNETITE: All black, ferromagnetic, opaque material is called magnetite (*see* Vincent and Phillips, 1954, p. 3). All of it is believed to contain some titania, and some, at least, is truly ilmenite.

CHLORITE: The chlorite has X = nearly colorless, Y = Z = pale green to blue-green. Anomalous blue interference colors are in the lower first order. These are the characters given for the variety penninite (Winchell, 1951).

ACCESSORY MINERALS FROM BANDED GNEISS: Accessory minerals were separated from a pair of adjacent felsic and mafic bands in the gneiss (G52a, G52b); bromoform was used to remove quartz and plagioclase, and a Franz Isodynamic Separator was used to separate, as much as was practical, the individual minerals. Each magnetic fraction was weighed on an analytic balance, quartered, and mounted in balsam. Grain counts then permitted calculation of the percentage of each accessory mineral in the total sample. Large samples were used, and extreme care was taken to minimize contamination. Table 5 gives the results of the separations.

Minerals in the felsic band include zircon, apatite, monazite, and allanite.

Zircon: colorless, zoning absent, but 20 per cent of individuals show outgrowths and embayments, indicating growth and/or partial resorption during one or more cycles of crystallization; mineral and fluid inclusions in all individuals. The only inclusions identified were euhedral apatite showing simple $(10\bar{1}0)$ and $(10\bar{1}1)$ forms. Zircon crystals are mostly simple (110) prisms with (111) terminations. A steeper dipyramid, probably (113), is present on many individuals; the rare pinacoid (001) occurs in at least one individual. Elongation (c/a) ranges from 2 to 9, with most individuals between 2 and 3.5.

Apatite: all individuals subhedral to rounded;

more than 50 per cent of the crystals are elongated parallel to c. Although no sharply euhedral crystals are present, the first-order prism, one or more dipyramids, and the pinacoid can be recognized in many individuals.

TABLE 5.—ACCESSORY MINERALS FROM SPECIMENS OF BANDED GNEISS

Felsic band		Mafic band	
Mineral	Percentage	Mineral	Percentage
Zircon	<0.1	Apatite	<0.01
Apatite	<0.1	Allanite	<0.001
Monazite	<0.01	Kyanite	<0.001
Allanite	<0.001	Rutile	<0.0001

Incipient basal parting (0001) is almost universal in elongated individuals: many of the fragments present are broken crystals lying on this face.

Acicular inclusions, opaque and transparent, are oriented parallel to the prism-zone axis in approximately half the crystals. Irregular opaque inclusions are present in a few individuals, and at least two apatite crystals contain inclusions recognizable as zircon.

Monazite: pale yellowish green and nonpleochroic; equant; crystal faces recognizable but not identifiable; well-developed basal pinacoid (001) probably common, because Bx_a) is vertical in about one-fourth of cases: $(+)2V < 20°$.

Allanite: brown, metaminct; birefringence, nil; $n = 1.72$; nonmagnetic.

Minerals in the mafic band include apatite, allanite, kyanite, and rutile.

Apatite: all grains irregular to subhedral; occasional prism faces $(10\bar{1}1)$ recognizable. Two types of inclusions occur. The first are acicular and opaque, oriented parallel to c. The second are very irregular anisotropic inclusions occuring in parallel trails showing no apparent relationship to the crystallographic directions of the host.

Allanite: as in felsic bands.

Kyanite: large platy grains, colorless and nonpleochroic. No inclusions; $(-)2V = 85°$ (estimated); $Z \wedge c = 32°$.

Rutile: Equant, subhedral grains. The crystal faces present were not identified. Pleochroic in shades of yellow and brown. Uniaxial, positive. Birefringence extreme.

BANDED GNEISS

OCCURRENCE: Bascom and Stose (1932, p. 6 considered the banded gneiss which makes u more than half the complex to be a highl silicic differentiate of a great gabbroic intrusion The vague term "banded gneiss" is used in th present account to avoid such terms as gabbro metagabbro, or migmatite, which may hav strong genetic implications.

On fresh surfaces the gneiss appears medium to fine-grained and quite massive. Veinlets o irregular masses of quartz, some containing large crystals of hornblende or biotite, occu sparingly.

Weathered surfaces expose planar and linea elements and particularly irregularly alter nating layers or bands having markedly differ ent proportions of felsic and mafic minerals The darker, more mafic bands are commonly 5–8 inches thick but may be nearly 5 feet thick (D4). Most thin bands are deformed and broken, whereas most thick ones persist for the entirety of an exposure, commonly several score feet. The mafic bands are spaced from foot to 12 feet apart.

The material between the dark bands has conspicuously coarser grain and lighter color because of the greater proportion of quartz and relative decrease in the amount of ferromag nesian minerals.

This lighter material intrudes openings and joints in the dark bands or may completely enclose isolated tabular pieces a foot or two long of the darker-phase material. The transition between the continuous dark bands and these detached remnants can be seen in Wilmington quarry (D4). Here three parallel mafic bands are exposed for about 20 feet. Two each about 2 feet thick, are continuous and show little sign of deformation. The third, narrower band, has been broken into angular pieces about a foot long and engulfed by the felsic material.

A distinctive variety of the banded gneiss exposed near the contact with the Wissahicko formation in the vicinity of Rockland, Dela ware. The gneissic banding cannot be detected at the immediate contact with the metased ments (G132) and first becomes visible about 15 feet from the contact. For abou half a mile from the contact to the first a pearance of typical gneiss with wide bands, t banding ranges in thickness from less than on quarter inch to about 4 inches. The bands tend to become thicker away from the contact un the typical gneiss is reached. Individual ban

in this portion of the gneiss have compositions and textures nearly identical with the typical material. However, the felsic bands comprise north of the New Castle County Workhouse the Wissahickon formation contains three sills of the felsic material (C24b).

TABLE 6.—MODES OF 10 ADJACENT PAIRS OF MAFIC AND FELSIC BANDS FROM BANDED GNEISS
(In volume per cent)

Mafic Bands

	C19b*	D4b	G52b	G101b	G125a	H2b	H19a	J1a	J6a	J9a	Range	Average
Plagioclase	58	36	64	45	55	43	47	32	27	48	27–64	45.5
An content	An 33	An 64	An 46	An 42	An 48	An 43	An 54	An 64	An 43	An 57	33–64	49
Quartz	2	..	2	..	..	8	..	10	18	16	0–18	5.6
Hornblende	37	17	6	43	..	tr	14	57	50	28	0–57	25.2
Hypersthene	..	24	16	7	31	38	29	..	..	6	0–38	15.1
Augite	1	22	10	..	8	6	8	..	..	..	0–21	5.4
Biotite	tr	..	tr	tr	tr	4	..	1	4	tr	0–4	1
Magnetite	2	1	1	5	6	2	2	tr	tr	2	0.2–6	2
Apatite	tr	..	tr	tr	tr	tr	tr	tr	tr	tr		
Zircon	tr	..	..	tr	..	tr	tr	..	tr	tr		

Felsic Bands

	C19a	D4a	G52a	G101a	G125b	H2a	H19b	J1b	J6b	J9b	Range	Average
Plagioclase	50	48	50	30	61	52	60	46	36	49	30–60	47.3
An content	An 30	An 44	An 37	An 40	An 44	An 42	An 50	An 46	An 44	An 48	30–50	43
Quartz	39	44	44	59	18	27	32	24	47	32	18–59	36.6
Hornblende	tr	..	..	..	..	tr	tr	1	1	1	0–6	0.8
Hypersthene	3	6	3	9	15	8	12	..	..	14	0–18	8
Augite	..	..	..	tr	tr	..	1	..	..	..	0–0.9	0.1
Biotite	7	tr	1	3	3	11	1	27	7	1	0.1–27	5.7
Magnetite	2	2	1	3	3	2	4	tr	2	3	0.3–4	2
Apatite	tr	tr	tr	tr	tr	tr	tr	tr	tr	tr		
Zircon	tr	tr	tr	tr	tr	tr	tr	..	tr	tr		
Orthoclase	..	..	..	..	..	tr	tr	..	..	..		

* Specimen locations shown on Plate 1

ly 30–50 per cent of this portion, in contrast o 60–80 per cent elsewhere. This material is ot treated separately in the section on compo- ition and texture. Five modes in Table 7 are om this portion of the mass (G11, G15, G98, 103, and G132).

At two exposures sills of a material identical ith the felsic phase of the gneiss intrude rocks ordering the banded gneiss. On the west side Brandywine Creek at Rockland, Delaware, ornblende-plagioclase amphibolite contains sills coarse-grained quartz-plagioclase rock (G5c) pparently identical with the light-phase rock. In the Red Clay Creek about 500 yards

Because of the distinct differences in the petrography of the mafic and felsic phases, they are treated separately.

MEGASCOPIC APPEARANCE: Hand specimens of the mafic bands are grayish-black or bluish black and so fine-grained that individual minerals can rarely be distinguished and identified with certainty. Gneissic structure usually cannot be detected in hand specimens, even though it is conspicuous on the weathered or wet surfaces of the outcrop, because the plagioclase is nearly as dark as the pyroxene. Near the contact with the Arden granite, the texture is coarser, and the individual grains can be

TABLE 7.—MODAL ANALYSES, BANDED GNEISS
(Modes in volume per cent)

	C24b*	D2a	D2b	G5c†	G11	G15	G47a	G47b	G98	G103	G132	35-5	35-6	35-8	35-9	35-19	28-74†
Plagioclase	30	53	60	28	52	24	50	60	57	59	45	59	50	64	60	57	50
An content	An 72	An 57	An 24	An 44	An 55	An 54	An 41	An 40	An 67	An 52	An 80	An 72	An 58	An 46	An 52	An 69	An 40
Quartz	35	43	39	55	14	65	32	30	11	12	4	..	13	27	..	..	38
Hornblende	4	..	..	..	10	5	8	tr	13	21	2	23	14	..	..	13	tr
Hypersthene	27	2	..	3	4	5	..	8	15	4	35	10	11	5	18	21	10
Augite	..	..	..	..	13	..	..	..	3	..	10	6	7	1	19	9	1
Biotite	tr	..	..	10	..	..	tr	tr	..	2	tr	..	..	tr	..	..	tr
Magnetite	3	2	tr	3	6	1	4	2	1	3	3	2	5	3	3	tr	1
Apatite	tr	tr	tr	tr	tr	tr	tr	tr	tr	tr	tr	tr	tr	tr	tr	..	tr
Zircon	tr	tr	tr	tr	tr	tr	tr	tr	..	tr	tr	..	..	..	..	..	tr
Epidote	..	..	..	..	..	..	..	..	..	..	..	..	..	..	tr	..	..
Tourmaline	..	..	..	tr	..	..	..	..	..	..	..	..	..	..	..	..	..
Chlorite	..	..	..	tr	..	..	tr	tr	..	..	..	..	..	..	..	..	..
Garnet	1	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..

* Specimen locations shown on Plate 1
† Estimated

distinguished easily. The increased grain size gives the rock the much lighter appearance that should be expected from rocks which contain approximately 50 per cent feldspar. In these coarser-grained hand specimens there is a lineation resulting from parallel disposition of hornblende grains and a gneissic structure resulting from a tendency for ferromagnesian minerals to be concentrated in narrow streaks or bands.

Hand specimens of the felsic bands typically appear dark blue gray and massive. The grain size, up to 2.5 mm, is distinctly greater than that of the associated mafic bands, but only the quartz and plagioclase can be distinguished and identified with confidence, because ferromagnesian grains are small and their color does not contrast with the bluish plagioclase and quartz. Near the contact with the Arden granite the grain is coarser, the color paler, and the component minerals more easily identified. Gneissic and schistose structures are more clearly defined.

COMPOSITION AND TEXTURE: About 75 thin sections of the banded gneiss have been examined. In addition, 10 representative outcrops scattered throughout the gneiss were selected for special study. Thin sections were prepared from adjacent felsic and mafic bands in each, and mineral descriptions and modal analyses were made. The percentages and ranges in composition in the following section are based upon these values (Table 6). The modes of other specimens to which reference is made are given in Table 7.

The mafic bands are made up largely of andesine or labradorite and one or more of the ferromagnesian minerals hypersthene, hornblende, and augite. Where biotite is present, it occurs in very small amounts as a late constituent. A little quartz occurs in most specimens, but only in the strongly sheared specimens at the northern border of the body does it reach a significant amount (J1a, 10 per cent; J9a, 16 per cent; J6a, 18 per cent). All samples contain minor quantities of magnetite and apatite.

The most distinctive petrographic characters of the mafic bands are the relatively fine grain and granoblastic texture. The plagioclase ranges from approximately 0.30 to 0.75 mm and the pyroxene from 0.10 to 0.75 mm. The plagioclase and pyroxene are totally anhedral, showing the mosaic texture characteristic of quartzofeldspathic gneisses and granulites. Nowhere has a palimpsest igneous or sedimentary texture been detected.

Where hornblende accompanies the pyroxenes it is clearly secondary and occurs as rims of small crystals on pyroxene grains; as complete, or nearly complete, replacements of pyroxene grains retaining the equant shape of the original grains; and as lath-shaped crystals cutting the pyroxene grains or masses. In some specimens hornblende replacement is concentrated in layers or bands several millimeters thick, parallel to the gneissic structure and the coarse banding. Where hornblende has replaced pyroxene extensively, the granoblastic texture is obscured by the slightly schistose texture, which results from the elongation of hornblende grains. With one exception, the textures indicate that the dark bands at one time were entirely plagioclase-hypersthene-augite: the pyroxene grains are not pseudomorphous after hornblende, nor can hornblende be seen in the process of alteration to pyroxene, whereas the opposite process can be observed in almost every slide. The exception is J1a, at the extreme northern end of the gneiss, where hornblende and biotite occur without pyroxene. This rock has been strongly sheared, and there is no conclusive evidence that the hornblende has been derived from earlier pyroxene.

Biotite, which is present only in very small quantities, is apparently the last mineral to have formed. It occurs as rims around magnetite, as irregular masses replacing pyroxene and hornblende, and rarely as large laths cutting hornblende grains.

Crushing, granulation, and other effects of dynamic forces contemporaneous with, or postdating, recrystallization are minor or absent in the bark bands. Many twin lamellae in plagioclase grains are bent, although none are broken, and quartz, where present, is severely strained, although rarely granulated. However, in areas where the dark bands have been deformed, broken, and partially digested by the associated felsic material, weak cataclastic texture is developed, and quartz may show some peripheral granulation.

The more abundant felsic material is made up largely of quartz and plagioclase, and some very coarse-grained felsic bands are composed almost entirely of quartz and plagioclase. One (D2a) contains 2 per cent each of magnetite and hypersthene. Another (D2b) contains, in addition to quartz and plagioclase, only 0.5 per cent magnetite.

The felsic phase, although granoblastic, lacks the even-grained, mosaic texture which is present in many mafic bands. The difference is due largely to the nature of the quartz, which has an irregular, lobate habit, embaying and replacing the surrounding plagioclase and

pyroxene. The quartz ranges in size from about 0.4 to 2.5 mm, whereas the associated feldspar and pyroxene rarely exceed 1 mm.

Hornblende is present in 60 per cent of the slides of the felsic phase, but in only one-fifth of them does it exceed 1 per cent. It is of late origin, and its effect upon the texture of the rock is small. Biotite, which is present in the felsic material of all paired samples, occurs as irregular replacements of pyroxene or hornblende or as laths cutting them. As in the dark bands, it appears to have been the last mineral to crystallize or recrystallize.

The pyroxene, dominantly hypersthene, occurs as small, equant, rounded grains concentrated in poorly defined streaks. Some groups of grains show identical optical orientation. Large irregular pyroxene grains may be only partially replaced by plagioclase and quartz. This appears to display an earlier stage in the reduction of larger pyroxene crystals to masses of small rounded grains.

The felsic material commonly shows evidence of considerable crushing. Quartz is severely strained and may be peripherally granulated. Plagioclase crystals have bent twin lamellae, pyroxene may be granulated, and larger biotite plates may be twisted and crumpled. Thin sections from a quarry on Brandywine Creek (G47a, G47b) reveal extreme cataclasis. Ovoid plagioclase and hypersthene are set in a fine-grained mylonite made up largely of quartz. Hypersthene is peripherally converted to small irregular hornblende shreds, and some hornblende is partially altered to chlorite.

Arden Granite

OCCURRENCE: Approximately 4 square miles in the vicinity of Arden, Delaware, is underlain by a potassium feldspar-bearing rock which in some places has the composition of true granite. It is bordered on three sides by the banded gneiss and on the east is covered by Quaternary and Recent sediments. Actual contacts with the surrounding rock are nowhere exposed, but in Naaman's Creek an outcrop of gneiss (H2) is separated from an outcrop of granite by only 50 yards. From the character of the two exposures some inferences can be made concerning the relationship of the two rock types. At H2 the foliation and banding are irregular, and evidence of mobilization is greater than at most exposures of the gneiss. The granite at H25 contains xenoliths about 6 inches long of partially digested mafic bands from the gneiss. The xenoliths, which display the cross joints typical of the mafic bands, are aligned parallel to the weak foliation of the granite. From this it appears that the granitic material is younger than the banded gneiss and probably was emplaced as a more or less mobile material. Mineralogically, considerable biotite has developed at the expense of earlier ferromagnesian minerals, giving evidence of potassium metasomatism. Pegmatoid segregations within the mass and satellite pegmatites such as are commonly associated with granites are absent.

MEGASCOPIC APPEARANCE: In hand specimen and at the outcrop the granite is medium- to fine-grained and greenish gray. Phenocrysts of vitreous potassium feldspar showing Carlsbad twinning reach lengths of up to 20 mm (H15, H15). At most exposures a weak foliation is apparent, with small flakes of biotite aligned in a common plane. Variations in composition are not easily detected in the outcrop, but examination in index oils of a series of chips taken at intervals of several feet along a Baltimore and Ohio railroad cut (H4) shows considerable variation in the nature of the ferromagnesian phases present. Biotite occurs alone or with hypersthene, augite, or hornblende. The last is quite rare, whereas orthopyroxene is present in every sizable hand specimen.

COMPOSITION AND TEXTURE: Table 8 gives the mineralogy of representative thin sections from the granitic mass. Two of these, H15a and H25, are granite in the strictest sense, comprising quartz, oligoclase, orthoclase, and biotite. The specimens of granodioritic composition, H15 and H25, differ from normal rocks of the granite clan in that they contain, in addition to biotite, hypersthene and augite. The most characteristic petrographic features of the rock are the porphyritic nature of the feldspars and the distinctly mylonized character. The specimens of granitic composition show most evidence of granulation, with the larger feldspar crystals set in a matrix of fine-grained mylonite composed of quartz-orthoclase-plagioclase or quartz-orthoclase-plagioclase-biotite. In the more mafic specimens crushing, although present, is much less pronounced.

The order of crystallization of the felsic minerals is not clearly indicated. Among the ferromagnesian phases, the biotite has crystallized later than the pyroxene, commonly at its expense. Potassium-bearing hydrothermal fluids are the probable agent of this transformation. Hornblende occurs only in small quantities and

appears to result from retrograde alteration of pyroxene at a lower temperature.

The orthoclase is invariably perthitic and the plagioclase antiperthitic. Oligoclase-quartz rocks with markedly schistose to nearly granoblastic texture.

The amphibolites of the complex occur as an apparently continuous body extending south-

TABLE 8.—MODAL ANALYSES, ARDEN GRANITE
(Modes in volume per cent)

	H4*	H15a	H15b	H20b†	H25	Average
Plagioclase	29	18	63	36	65	42
An content	An 16	An 15	An 26	An 26	An 30	An 23
Potassium feldspar	39	46	3	25	8	24
Quartz	29	3	9	39	17	18
Myrmekite	1	1	..	1	1	1
Biotite	1	5	3	4	3	3
Hypersthene	..	..	13	3	5	4
Augite	..	..	3	tr	tr	1
Hornblende	..	..	tr	..	..	
Magnetite	tr	tr	5	1	1	2
Apatite	..	tr	1	tr	tr	
Zircon	tr	tr	tr	..	tr	

* Specimen locations shown on Plate 1
† Estimated

myrmekite is common. In specimen H22 a group of small euhedral crystals of green spinel embedded in plagioclase preserves the outline of what was probably a lath of hypersthene.

Within the Arden granite a nearly complete series demonstrates that potassium, sodium, and silica entering along permeable crushed zones transformed a rock similar to the banded gneiss into a granite.

First phase (example, specimen H15b): crushing minor; plagioclase near andesine, An_{35}; orthoclase present only in antiperthite; peripheral conversion of pyroxene to biotite.

Second phase (examples, specimens H20b, H25): crushing marked; plagioclase unchanged, An_{26}, An_{30}; antiperthite common; orthoclase < plagioclase; biotite is the dominant ferromagnesian phase.

Final phase (examples, specimens H4, H15a): crushing extreme in places; plagioclase more sodic, An_{15}, An_{16}; orthoclase > plagioclase; biotite is the sole ferromagnesian phase.

AMPHIBOLITES OF THE COMPLEX

OCCURRENCE: Amphibolite is used here as a lithologic term to designate hornblende- plagioclase or hornblende-augite-plagioclase-epidote

westward from the banded gneiss near Wilmington for about 10 miles to northeastern Cecil County, Maryland. Their contact with the gneiss appears to be gradational in that the mafic bands in gneiss near the amphibolite are hornblende-rich, whereas the more easterly amphibolites contain a little hypersthene. The difference lies in the presence or absence of the felsic phase. The felsic phase of the gneiss, however, displays no systematic gradual decrease in quantity as the amphibolite is approached.

The contact with the Wissahickon formation, which extends for about 16 miles, is sharp and accordant with regional strike and with the strike of the foliation in both the schist and the amphibolite.

About half a mile from the contact with the Wissahickon formation, near Rockland, Delaware, there is a group of exposures extending along the west bank of Brandywine Creek, a Reading Railroad spur about 100 feet above the Creek, and along a road about 40 feet above the railroad. In these outcrops (F106, F107) thick beds of sillimanite-almandine-biotite gneiss (or schist) are interbedded with the amphibolite. The pelite is coarse-grained with conspicuous bedding which is defined in places by persistent thin layers composed entirely of quartz and oligoclase. The bedding and schis-

tosity are megascopically parallel to each other and to the regional strike. As nearly as can be determined, the pelitic rocks occur in two (or perhaps three) distinct layers, each about 40 feet thick, separated by about 10 feet of amphibolite. These beds can be traced laterally for about one-quarter of a mile and appear to grade laterally into amphibolite, but data are inconclusive. No evidence of faulting or of magmatic action such as would be expected in a xenolith exists. These beds can almost certainly be correlated with the nearby Wissahickon formation, which they resemble in both composition and texture.

Most of the southern boundary of the amphibolite is formed by the overlap of Cretaceous and younger sediments. Unravelling the structure and history of the amphibolites is difficult, because a large portion of the area lies along the Fall Line, where exposures are small and discontinuous.

The Port Deposit granodiorite forms the southern boundary of the amphibolites at the extreme western end of the outcrop belt. The amphibolite shows effects of the emplacement of the granodiorite, in that many exposures from Newark to Milltown, Delaware, contain small veins of microcline granite apparently related to the larger body. In the bed of White Clay Creek near Newark there is a relatively large outcrop (B52) where microcline granite has intruded the amphibolite, producing alternating bands of granite and amphibolite structurally similar to the banded gneiss. Bright-green epidote is developed at the contact between the bands. Pod-shaped pegmatite bodies cut both rock types.

A distinctive and atypical variety of amphibolite is exposed in a large railroad cut about 2 miles east of Newark. The rock appears massive, bluish, and fine-grained (B33a). Weathered surfaces reveal weak linear and planar elements. Megascopically it is similar to the mafic phase of the banded gneiss, but nothing resembling the felsic phase is exposed. There are occasional sinuous darker bands (B33b) several inches thick in the already quite mafic material. Because no other outcrops occur nearby, the geologic relationships of this material with the other amphibolites are obscure.

MEGASCOPIC APPEARANCE: In hand specimens the amphibolites are medium- to fine-grained, dark rocks, typically displaying a weak planar structure defined by varying proportions of hornblende, plagioclase, and quartz in adjacent layers or streaks. Lineation may be marked, where hornblende crystals are large, or nearly absent in the finer-grained specimens. The quartz and plagioclase lack the blue color which they have in the banded gneiss. Pyroxene, where present, cannot be distinguished with certainty.

The fine-grained plagioclase-hornblende rock has abundant spherical to ovoid areas up to cm in diameter comprising quartz and plagioclase, commonly with a central porphyroblast of magnetite. Units having this texture range from several inches to a few feet in thickness and have been discovered in outcrops along a belt stretching from Brandywine Springs to Rockland, Delaware. Where present, the "spots" make up 25–50 per cent of the rock. They have a remarkable resemblance to amygdaloidal texture. Their mineralogy is discussed in the following section.

COMPOSITION AND TEXTURE: The amphibolites of the complex may be divided into two compositional groups.

Group *A* specimens (Table 9) are concentrated in the eastern portion of the amphibolite area near the banded gneiss. The texture ranges from granoblastic to schistose. Plagioclase and hornblende predominate; a little hypersthene is present but is markedly out of equilibrium and partly converted to hornblende. In the more granoblastic specimens the equant hornblende grains appear to be, in part, pseudomorphous after hypersthene. The hornblende is pleochroic yellow and green, occasionally showing the olive cast of that in the banded gneiss. Quartz and biotite are absent or inconspicuous, and crushing is minor. As in the banded gneiss, plagioclase at least as calcic as An_{72} may exist and is not accompanied by epidote. Biotite is inconspicuous or absent in most specimens. There is, however, a group of relatively K_2O-rich specimens concentrated near the contact with the Wissahickon formation (F22, F47, F48, F63). In these, the amount of brown biotite equals or exceeds that of hornblende, and in one instance (F22) potassium was sufficiently abundant to permit crystallization of orthoclase. The most potassic specimens also appear richer in sodium than the other amphibolites, having plagioclase as sodic as An_{25-30}.

The grain size and texture, which vary widely within a few yards, appear to be related to the nature of deformation and directional pressures acting in any small area as well as to the composition and texture of the premetamorphic rock.

The amygdaloids (?) present in some specimens of this group consist of hornblende

TABLE 9.—MODAL ANALYSES, GROUP *A* AMPHIBOLITES
(Modes in volume per cent)

	B50*	B51	C18a	C18b	C20a	C20b	D1	F72	F75	F138	G56	J2
Plagioclase	25	42	61	44	33	30	50	55	26	46	54	45
An content	An 42	An 52	An 46	An 56	An 44	An 53	An 72	An 71	An 68	An 72	An 70	An 68
Hornblende	75	50	30	40	45	20	47	40	60	40	44	35
Quartz	..	5	8	16	17	40	tr	4	5	10	..	20
Hypersthene	..	3	..	..	..	..	..	1	6	4	..	..
Biotite	tr	..	tr	tr	1	5	..	..	..	..	..	..
Opaque	tr	..	1	tr	3	5	3	tr	3	tr	2	2
Apatite	..	..	tr	tr	tr	tr	..	tr	..	..	..	..
Zircon	..	..	..	..	tr	tr	..	tr	..	tr	..	..
Allanite	..	..	..	..	..	tr	..	..	..	..	..	..

*Specimen locations shown on Plate 1

TABLE 10.—MODAL ANALYSES, GROUP *B* AMPHIBOLITES
(Modes in volume per cent)

	A60*	B29a	B29b	B33a	B33b	B52a
Plagioclase	43	30	40	45	48	51
An content	An 36	An 42	An 44	An 73	An 69	An 24
Hornblende	15	..	50	27	32	43
Augite	..	..	..	27	19	..
Biotite	..	25	tr	..	..	..
Epidote	6	2	1	1	1	4
Garnet	tr	..	..	..	..	..
Quartz	30	40	8	..	..	..
Opaque	2	2	tr	tr	tr	tr
Chlorite	..	..	..	..	..	1
Zircon	..	tr	tr	..	..	tr
Apatite	tr	tr	tr	tr	tr	tr
Sphene	..	tr	1	..	..	1
Rutile	..	tr	tr	tr	..	..

*Specimen locations shown on Plate 1

areas in otherwise hornblende-rich rock. Their structure is concentric in that the magnetite porphyroblast is central. The plagioclase and quartz, in some places with a trace of biotite, which make up the remainder of each "spot" are granoblastic and do not have radiating structure; the plagioclase has the same composition as the remainder of the rock. In the metavolcanic rocks of Cecil County, Maryland (Marshall, 1937), which are about 10 miles from the southwestern end of the Wilmington complex, amygdaloidal texture is widespread. The amygdules are described by Marshall (1937, p. 195–199) as being made up of plagioclase (An_{20}–An_{30}), epidote (which partially replaces the plagioclase), quartz, and magnetite. In some of Marshall's specimens orthoclase is reported.

Excepting the orthoclase-bearing variety, the amygdaloids of Cecil County appear to be very similar chemically and mineralogically to the "spotted" amphibolites of this area; the substitution of An_{20-30} and epidote for more calcic plagioclase reflects a difference in the conditions attendant upon recrystallization rather than a difference in composition. In light of this similarity, and in the absence of a rival theory for explaining the existence of this unusual texture, the possibility that this is a relic amygdaloidal texture must be considered.

Group *B* amphibolites (Table 10) make up the extreme western end of the complex. This

category includes a diverse group of rocks whose common characteristic is the partial replacement of calcic plagioclase by epidote. Hornblende may be the sole ferromagnesian phase or may be replaced by the pair diopside-

(Bascom and others, 1909; Bascom and Miller, 1920; Bascom and Stose, 1932). In conjunction with the present study, several of these bodies within the Wissahickon formation near the complex were examined and sampled. The in-

TABLE 11.—MODAL ANALYSES, GROUP *C* AMPHIBOLITES

(Modes in volume per cent)

	A54*	E25	E61a	F2	F19	F54
Plagioclase	20	30	35	31	40	36
An content	An 42	An 46	An 44	An 78	An 62	An 80
Hornblende	70	60	60	69	60	64
Quartz	5	8	..	..	..	..
Opaque	3	2	5	tr	tr	tr
Biotite	1	..	tr	..	..	..
Zircon	..	tr	tr	..	..	tr
Apatite	..	tr	tr	..	..	tr

* Specimen locations shown on Plate 1

hornblende (B33a, B33b). In one exposure (A60) almandine-hornblende is present. These last-mentioned mineralogical differences can be attributed to differences in bulk composition, whereas the differences between the plagioclases of groups *A* and *B* suggest differences in their metamorphic history.

Hypersthene is absent in these rocks. Excepting B33, the specimens are schistose, quartz and biotite are abundant, and crushing is conspicuous. The hornblende is pleochroic in shades of bluish green. Specimen B33 is massive and granoblastic; later biotite and quartz are absent, and hornblende and diopside appear identical with those of the banded gneiss.

Where the amphibolite has been intruded by the Port Deposit granodiorite, a zone a few inches thick (B52) adjacent to the granodiorite contains microcline and abundant epidote. Hornblende has been partially converted to chlorite and the plagioclase partly converted to clay minerals. About 100 feet away a small zone exists in which biotite is the only ferromagnesian mineral present (B29a). This may represent complete alteration of hornblende by alkali solutions derived from the granodiorite.

AMPHIBOLITES OF THE WISSAHICKON FORMATION

OCCURRENCE: Within the Glenarm series and the Baltimore gneiss, which underlie much of the central Piedmont, previous workers have mapped large numbers of "gabbro bodies"

formation gained is of sufficient importance for the present study to warrant inclusion here, even though these bodies are not contiguous with the main complex.

Those amphibolite bodies contained entirely within the Wissahickon formation appear on Bascom's maps as narrow dikes and irregular bodies ranging from about 100 yards to more than 2 miles long. For some of these bodies no bedrock exposures exist. During the present study many new exposures were examined along with the older ones. All contacts seen were concordant. Furthermore, recent roadcuts show that what had appeared to be large irregular bodies are actually groups of smaller concordant ones. There are, unfortunately, many places for which no exposures exist, but in the absence of evidence to the contrary it seems better to consider these amphibolites to be concordant bodies within the Wissahickon formation rather than irregular plutons. Large ones may be more than 100 feet thick, but exposures are so small and widely separated that areas in which only the more resistant amphibolite is exposed may in fact be underlain largely by schist.

COMPOSITION AND TEXTURE: The mafic bodies within the schist fall into two distinct categories which closely resemble in distribution and composition the two groups present within the complex.

Group *C* (Table 11) comprises simple hornblende-plagioclase bodies with schistose to nearly granoblastic texture which resemble the

group *A* amphibolites. Crushing is absent, and quartz and biotite are either absent or minor. No hypersthene has been detected. The plagioclase ranges in composition from An_{44} to An_{80} and is not accompanied by epidote. These specimens are found within about 5 miles of the complex.

TABLE 12.—MODAL ANALYSES, GROUP D AMPHIBOLITES

(Modes (estimated) in volume per cent)

	B15*	E19	F70	F85
Plagioclase	30	45	28	45
An content	An 72	An 50	An 39	An 60
Hornblende	45	40	30	5
Quartz	1	..	..	..
Augite	20	7	20	43
Epidote	3	7	12	5
Opaque	tr	tr	tr	2
Sphene	1	tr	3	..
Calcite	..	..	2	..
Scapolite	..	..	5	..
Apatite	..	..	..	tr
Zircon	..	tr	..	..

*Specimen locations shown on Plate 1

Group *D* is represented by four specimens (Table 12) which occur at a greater distance from the contact with the complex. These are medium- to fine-grained, faintly gneissic rocks, containing plagioclase (An_{39}-An_{72}), epidote, clinopyroxene, and hornblende. Appreciable quantities of sphene are present in all, and one (F70) contains scapolite and calcite. The texture is granoblastic and mildly cataclastic. The bright green, weakly pleochroic clinopyroxene differs in appearance from the nearly colorless diopside of the banded gneiss. The hornblende appears in places to replace the clinopyroxene, but the effects of cataclastic action obscure this relationship. The epidote is of late origin, having developed largely at the expense of the feldspar, of which almost every grain is zoned, having a small, more calcic core.

GABBRO

OCCURRENCE: Between Newark, Delaware, and Elkton, Maryland, there are three small areas, ranging from about half a mile to 2 miles across, which are underlain by gabbroic and ultrabasic igneous rocks, including serpentine. Another body, 600 yards long and 350 yards wide, intrudes the banded gneiss at Bringhurst Park on the northern outskirts of Wilmington.

The contact relationships of the Iron Hill stocks with each other and with the country rock are obscure, because they occur as a group of low hills surrounded by Coastal Plain sediments. On the basis of very few data, mostly drill samples from water wells, it appears that the entire group is surrounded by the Port Deposit granodiorite, and that the two smaller bodies are separated from each other and from the larger body by the granodiorite. No exposures have been found showing inclusions or apophyses of either rock type in the other.

At Bringhurst the contact with the gneiss is well exposed. Several xenoliths of dark bands from the gneiss are enclosed in the gabbro within about 200 feet of the contact. These parallel the regional strike. Their orientation appears to be relict, rather than superimposed, for the host rock is massive and shows no sign of postcrystalline deformation.

MEGASCOPIC APPEARANCE: Fresh specimens from the Iron Hill stocks are dark and coarse- to extremely coarse-grained (0.5 cm to 7 cm). Feldspar-bearing specimens show least weathering, whereas the serpentines, apparently derived from peridotite, are deeply weathered and friable except where thoroughly silicified.

The Bringhurst stock is remarkably homogeneous and massive. All hand specimens contain nearly equal parts of plagioclase and pyroxene. The grain size ranges from about 3 cm in the north to about 1 cm at the southern end of the body. A macro-ophitic texture is clearly displayed at the more northerly end of the body but is less prominent in the finer-grained rock to the south.

COMPOSITION AND TEXTURE: The Iron Hill stocks display, strikingly, the process by which primary pyroxenes become converted to more or less fibrous amphiboles (uralitization). Chester (1890, p. 22–27) has given a detailed account of this uralitization.

The texture is hypautomorphic-granular. Plagioclase (An_{65}-An_{90}) ranges from 50 per cent of the total to less than 5 per cent. In those rocks with most plagioclase, uralitization of the pyroxene is complete or nearly so; some masses of secondary hornblende still contain small cores of hypersthene. The pyroxenites contain both orthopyroxenes and clinopyroxenes with little or no plagioclase. Here uralitization is less advanced or absent.

No unaltered olivine was seen, but many slides contain masses of serpentine which appear

to be pseudomorphs after olivine. The former existence of peridotite in the more westerly stock is indicated by the presence, at the surface, of siliceous material which in thin section is made up of cryptocrystalline silica containing minor serpentine and chromite. This material preserves in detail the shape and texture of the olivine which was originally the major constituent of the rock.

In the gabbroic specimens zoisite is associated with the plagioclase, either as large crystals or as small acicular inclusions. The plagioclase, nevertheless, is extremely calcic. A chemical analysis by Chester (1890, p. 28) gives a composition of An_{92}.

The Bringhurst stock north of Wilmington comprises subequal portions of plagioclase (An_{60}) and ferromagnesian minerals.

The plagioclase shows little zoning and is not associated with epidote minerals. Opaque inclusions having stubby prismatic form are present, along with a few small euhedral crystals of what appears to be clinopyroxene.

The ferromagnesian minerals comprise hypersthene and augite with small quantities of later hornblende and biotite. Peripheral conversion of hypersthene to hornblende is most common where the hypersthene is in contact with plagioclase, which provides a source of Al_2O_3 and CaO. In the augite, conversion to amphibole is less dependent upon external sources of Al_2O_3 and CaO, and the process is more widespread. Schiller structure, which is nearly universal in the pyroxenes, results from oriented inclusions of two types: (1) square plates of a black, opaque mineral, presumably magnetite; and (2) brown, translucent plates with micaceous habit, possibly hematite.

The very minor biotite occurs as a replacement of hornblende. Whether it is a product of the end-stage fluids of the gabbro magma or whether it resulted from the postcrystalline potassium metasomatism which affected the country rock could not be determined.

The xenoliths in the gabbro are typical dark bands from the banded gneiss in composition and texture. A significant feature of the xenoliths is that the pyroxenes have the same schiller structure as the pyroxenes of the host rock. Inasmuch as this does not exist elsewhere in the gneiss, it must be a result of the action of the magma upon the zenolith. The magma reacted with the pyroxenes of the xenolith to convert them to compositions with which the magma was saturated and which were crystallizing, or about to crystallize, at that time. The cooling and postcooling histories of the pyroxenes of both rocks are identical, as are the products of the process (Bowen, 1928, pp. 175–201).

TABLE 13.—MODAL ANALYSES, PORT DEPOSIT GRANODIORITE

(Modes (estimated) in volume per cent)

	A47*	A102	B52b	28-53	28-36
Plagioclase	15	40	56	40	40
An content	An 20	An 32	An 15	An 22	An 24
Quartz	40	35	33	35	30
Microcline	40	10	7	20	25
Biotite	2	10	..	4	4
Magnetite	tr	tr	1	1	tr
Epidote	tr	6	1	..	tr
Apatite	tr	..	..	tr	tr
Zircon	tr	tr	1	tr	tr
Garnet	1	2	..	..	tr
Muscovite	1	1	..	..	1
Chlorite	..	..	2	tr	tr
Sphene	..	tr	..	..	..

* Specimen locations shown on Plate 1

PORT DEPOSIT GRANODIORITE

OCCURRENCE: This body underlies a large area southeast of the complex (Hershey, 1937). It apparently surrounds the Iron Hill stock and has intruded, *lit-par-lit* or in veins, the amphibolites along the southern edge of the complex. Several small bedrock exposures occur near the Delaware-Maryland boundary, between Iron Hill and the westernmost amphibolites. Deeply weathered specimens of what appears to be the same material have been taken from drill holes in the Coastal Plain sediments around the Iron Hill stocks. Weathered bedrock from the bottom of deep wells (800 feet) as far south and east as Delaware City resembles the granodiorite.

MEGASCOPIC APPEARANCE: Hand specimens of the veins and dikes intrusive into the amphibolite are relatively fine-grained. The rock appears greenish or pinkish, depending on the color of the microcline. A distinct foliation is defined by individual flakes of biotite which are aligned in a common plane. No gneissic segregation is discernible. Specimens from the main mass of the granodiorite are coarse-grained, light-colored rock with lenticular dark knots comprising biotite and epidote. These specimens appear identical with one

from type exposures along the Susquehanna River.

COMPOSITION AND TEXTURE: Modes of five specimens are given in Table 13. The microscopic texture of these rocks is xenomorphic-granular but has been somewhat modified by deformation and recrystallization. The quartz is but slightly strained. Subhedral plagioclase (oligoclase) shows hydrothermal alteration. The microcline is fresh, interstitial, and ranges from about 15 per cent to nearly 50 per cent of the total. The biotite, pleochroic in shades of brown or green, is associated with epidote and a few small euhedral garnets. The green biotite shows peripheral alteration to chlorite. In B52b, where the plagioclase shows extreme hydrothermal alteration, chlorite has completely replaced biotite.

METAMORPHIC FACIES

General Statement

The rock types previously described are classified according to the metamorphic-facies system proposed by Eskola (1939) and elaborated upon by Turner (Turner, 1948; Turner and Verhoogen, 1951; Fyfe, Turner, and Verhoogen, 1958). This system is better adapted to the study of mafic rocks than the Barrovian zone system (Barrow, 1893), which has been widely used for classifying pelitic rocks.

The facies represented here are those found in the medium and higher grades of regional metamorphism: granulite (or charnockite) facies and almandine amphibolite facies, including several subfacies within that group.

Banded Gneiss

The mafic bands of the gneiss are intensely crystalloblastic and appear to represent complete recrystallization. Whatever the origin of their chemical composition, the mineralogy is a result of the conditions of recrystallization.

The assemblage plagioclase-hypersthene-augite which characterizes the mafic bands is an equilibrium assemblage in the granulite facies, which represents highest-grade regional metamorphism (Eskola, 1939, p. 360; Turner, 1948, p. 100–104). Three chemical analyses of mafic bands are plotted on an ACF diagram (Fig. 4) showing mineral assemblages for the granulite facies (Turner, 1948, p. 101). Excepting the hornblende, which gives textural evidence of late origin, the mineralogy of the specimens is in rather good agreement with the predictions of the diagram.

The hornblende present in nearly all specimens of the banded gneiss has crystallized later than, usually replacing, pyroxene and is therefore a result of retrograde metamorphism occurring after equilibrium at the granulite-facies level was reached. Such diaphthoresis is widespread in rocks of extreme metamorphic grade (Turner, 1948, p. 103).

Some of the accessory minerals of the mafic bands give additional evidence of the conditions of crystallization. The rutile indicates metamorphism at or near the level of the granulite facies (Turner, 1948. p. 88, 102). Kyanite is also an indication of high-grade metamorphic conditions, although not necessarily higher than the upper portion of the almandine amphibolite facies. Its presence in a rock of this composition, however, warrants comment. The kyanite, which occurs as large platy crystals free from inclusions, was detected only in crushed material and not in thin sections, so there are no textural clues as to its origin. The facts that it is a stable phase at the high grade of metamorphism which the total mineral assemblage of the banded gneiss appears to represent and that kyanite is known in amphibolites which are chemically similar to the mafic bands here (Tilley, 1937) suggest that its presence in this relatively low-alumina rock is a result of a local chemical inequilibrium. Slides in the Bryn Mawr College collection made from an amphibolite within the Wissahickon formation half a mile west of Media, Pennsylvania (33–301), show small kyanite crystals formed along the contact between hornblende and a plagioclase grain which embays it. Apparently very locally the environment became sufficiently Al_2O_3-rich during metasomatism to permit the formation of an aluminous phase which could not be predicted or stable on the basis of the composition of the entire rock. This specimen of the gneiss, then, may represent a similar condition. Replacement of a relatively aluminous mineral such as plagioclase by quartz might result in a local Al_2O_3 excess, permitting kyanite to crystallize.

At the northern end of the banded gneiss adjacent to the Wissahickon formation there are small areas (J1, J6) where the mafic bands contain no pyroxene, only hornblende-plagioclase(-quartz-biotite). The hornblende is not pseudomorphous after pyroxene, so it is not possible to determine whether these bands are retrograde or whether they have never been at a higher grade than they now show. Never-

theless, it seems proper to assign this group of mafic bands to the almandine amphibolite facies and the remainder to the granulite facies.

The felsic phase of the gneiss is not so markedly crystalloblastic as the mafic bands. Foliation is also less pronounced than in the mafic bands. The most convincing evidence for metamorphic origin of the present mineral assemblage lies in the character of the clinopyroxene and of the hornblende, which are considered metamorphic in origin. Both of these, and the hypersthene, are idential, chemically and optically, with those of the mafic bands. Complete recrystallization of the felsic bands seems probable.

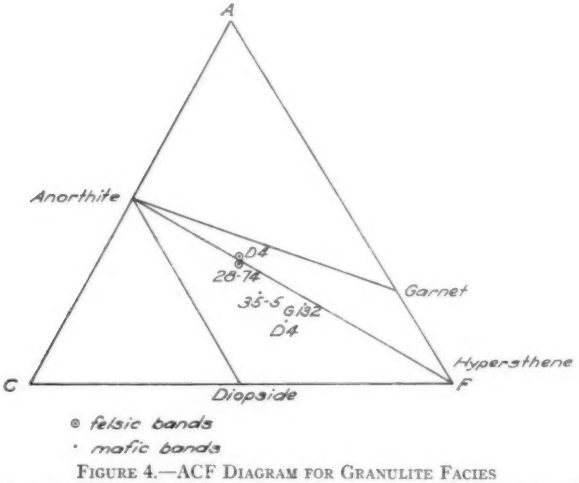

FIGURE 4.—ACF DIAGRAM FOR GRANULITE FACIES
$A = Al_2O_3 + Fe_2O_3 - (Na_2O + K_2O)$, $C = CaO$, $F = FeO + MgO + MnO$ (after Turner, 1948)

Two chemical analyses of felsic bands are plotted on an ACF diagram for the granulite facies (Fig. 4). The agreement is good, except that a little almandine might be expected in one specimen. Failure of this phase to appear may be a result of the proximity of the point to the hypersthene-plagioclase join.

Whether the present mineralogy is a result of igneous or metamorphic processes, the assemblage certainly indicates that the felsic material crystallized at the granulite-facies level wherever it occurs. Even in those places where the mafic bands contain no pyroxene, the felsic material does, and, moreover, the ratio of pyroxene to hornblende is higher in the felsic material than in the dark bands.

Arden Granite

Whether the processes by which this granite was formed are properly considered igneous metamorphic, it has characteristics which indicate crystallization or recrystallization conditions typical of the granulite (or charnockite) facies (Turner, 1948; Groves, 1935) including:

(1) the presence of the pair hypersthene augite—these phases are identical with the pyroxene of the banded gneiss;

(2) the presence of unaltered spinel in a rock containing excess silica;

(3) the presence of optically monoclinic potassium feldspar with small optic angle.

The character of the potassium feldspar may be significant for the history of both the granite and the complex as a whole. Recent studies of the potassium feldspars (Goldsmith and Laves, 1954) have demonstrated that the ordered triclinic form, microcline, is the truly stable phase at lower temperatures and that orthoclase is actually a metastable transition between monoclinic sanidine and triclinic microcline. X-ray-powder photographs (Heier, 1957) of potassium feldspar from several rocks representing different metamorphic facies demonstrate that the degree of inversion to triclinic form is least in the granulite and upper almandine amphibolite facies rocks and that the "orthoclase" representing this slight inversion optically appears monoclinic, or nearly so.

The presence in this granite of apparent monoclinic potassium feldspar and the complete absence of microcline indicate formation at temperature and pressure conditions equi

ent to the granulite facies, whereas the occurrence of individual crystals showing close similarity to sanidine—the true monoclinic form—indicates that the decline from high-temperature conditions was rapid and that processes which would have accelerated inversion were slight or absent. Reports of sanidine-like potassium feldspar in plutonic rocks are rare; orthoclase with an optic angle as small as 36° has been reported from charnockitic rocks of the Santa Lucia Mountains, California (Compton, 1957), although most grains measured by him showed larger values (Compton, personal communication). On the basis of the crystallographic work cited it seems likely that occasional crystals with a small optic angle should be found wherever postcrystallization metamorphism has been slight.

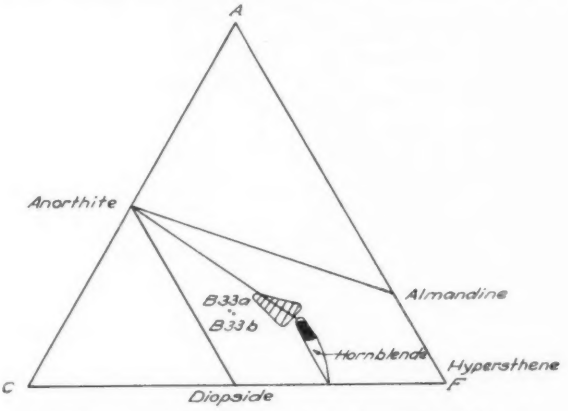

FIGURE 5.—ACF DIAGRAM FOR UPPER AMPHIBOLITE FACIES
$A = Al_2O_3 + Fe_2O_3 - (Na_2O + K_2O)$, $C = CaO$, $F = FeO + MgO + MnO$ (after Turner, 1948). Shaded area shows composition of analyzed hornblende; hatched area shows composition of mafic bands.

Amphibolite

The amphibolites of this report all clearly represent equilibrium or near-equilibrium assemblages within the almandine amphibolite facies (Eskola, 1939). This facies, however, covers a wide range of metamorphic conditions; it appears possible to make tentative subfacies assignments on the basis of the mineralogy.

Groups *A* and *C* are simple hornblende-plagioclase rocks, commonly with some quartz and a little biotite, which occur at or near the borders of the banded gneiss (Fig. 2). Hypersthene, markedly out of equilibrium and partly converted to hornblende, is present in some. The specimens which contain hypersthene are concentrated near the higher-grade banded gneiss and may show only weak schistosity as a result of the more or less equant shape of the hornblende grains, which may in part be pseudomorphs after hypersthene. Their position and texture and the lack of equilibrium between hornblende and hypersthene indicate that they are, in part, at least, the produce of diaphthoresis of a higher-grade assemblage. The absence of almandine or diopside from these amphibolites probably results from their bulk chemical composition. No chemical analyses of these rocks are available, but if these amphibolites are the less metamorphosed equivalents of the mafic bands in the gneiss, then the simple mineralogy results from the fact that their compositions lie on or very near the hornblende-plagioclase join in an ACF diagram (Fig. 5).

Although the assemblage hornblende-plagioclase is critical for all subfacies of the almandine amphibolite facies (Turner, 1948; Fyfe, Turner, and Verhoogen, 1958), it seems most likely that these rocks represent the sillimanite-almandine subfacies, for the following reasons:

(1) the very calcic nature of the plagioclase, which is not accompanied by epidote;

(2) proximity to the higher-grade banded gneiss;

(3) the fact that the Wissahickon formation which adjoins or encloses these amphibolites contains sillimanite-almandine and lacks lower-grade phases such as kyanite and staurolite.

The amphibolites of groups *B* and *D* (Tables 10, 12), which occur at greater distances from the banded gneiss than groups *A* and *C* (Fig. 2), are characterized by the presence of epidote

TABLE 14.—CHEMICAL ANALYSES OF ROCKS

	G132*	35-5	D4a	D4b	B33a	B33b	28-74	*X*	*Y*
SiO_2	52.14	45.85	69.19	48.99	50.25	50.50	67.26	67.67	46.49
Al_2O_3	15.60	18.55	14.44	14.61	16.87	16.64	15.04	16.87	17.73
Fe_2O_3	0.23	3.62	1.89	1.79	0.84	0.76	2.13	2.22	3.66
FeO	10.84	9.34	3.34	8.62	3.42	3.53	3.53	2.18	6.17
MgO	9.27	6.72	1.89	10.89	9.39	9.35	1.74	1.09	8.86
CaO	8.78	11.96	4.71	12.58	16.93	16.70	5.36	4.42	11.48
Na_2O	0.55	1.64	3.09	1.36	1.23	1.31	3.29	4.18	2.16
K_2O	0.04	0.16	0.39	0.15	0.10	0.07	0.35	0.57	0.78
H_2O+	0.27	0.35	0.21	0.26	0.46	0.56	0.30	0.29	1.04
H_2O-	0.04	0.00	0.04	0.03	0.00	0.00	0.14	tr	
TiO_2	1.40	1.08	0.55	0.34	0.13	0.13	0.41	0.28	1.17
P_2O_5	0.13	0.12	0.10	0.02	0.00	0.01	0.20	0.15	0.29
MnO	0.16	0.21	0.08	0.20	0.11	0.12	0.24	0.06	0.17
Total	99.45	99.60	99.92	99.84	99.73	99.68	99.99	100.05	100.00

* Specimen locations shown on Plate 1

G132 Mafic phase of banded gneiss 1 foot from contact with Wissahickon schist (analyst, C. Oliver Ingamells)

35-5 Mafic band in banded gneiss 4 miles north of Wilmington. Chemical analyses exist for hornblende, hypersthene, and augite from this specimen (analyst, C. Oliver Ingamells)

D4a Felsic band of banded gneiss from quarry in Wilmington (analyst, C. Oliver Ingamells)

D4b Mafic band adjacent to D4a (analyst, C. Oliver Ingamells)

B33a Amphibolite 1 mile east of Newark, Delaware (analyst, C. Oliver Ingamells)

B33b As B33a, but rock has slightly darker appearance (analyst, C. Oliver Ingamells)

28-74 Felsic band in banded gneiss 3 miles north of Wilmington (analyst, A. Meier)

X Trondhjemite; Malletuen, Norway (Johannsen, 1932, p. 385)

Y Average of 17 olivine gabbros (Johannsen, 1937, p. 222). Values for olivine-norites and olivine basalts are closely comparable with values for olivine gabbro

along with the plagioclase, which in most specimens is andesine, although more calcic plagioclase does occur. In many specimens augite accompanies hornblende as a result of a relative richness in CaO (Fig. 5). These assemblages may occur in rocks of appropriate composition throughout the almandine amphibolite facies and also in the hornblende granulite subfacies of the granulite facies (Fyfe, Turner, and Verhoogen, 1958, p. 235). This last subfacies formerly occupied an indeterminate position in the amphibolite facies (Turner, 1948). The presence of andesine-epidote in place of more calcic plagioclase indicates that these rocks were recrystallized at lower temperatures, probably with concomitantly higher stress, than the amphibolites of groups *A* and *C*. It may be noteworthy that these relationships are characteristic of the environment which has elsewhere produced the almandine-diopside-hornblende subfacies (Turner, 1939, p. 584–585)—now the hornblende-granulite subfacies.

Although the andesine-epidote in these rocks (and in the former almandine-diopside-hornblende subfacies) might indicate a relationship with the lower portion of the almandine amphibolite facies, a further complication occurs in the calcite-augite-hornblende-andesine-epidote-scapolite-quartz rock (F70) from group *D*. Here the absence of grossularite from a calcite-bearing rock also indicates kinship with the almandine-diopside-hornblende subfacies but tends to support the transfer of the subfacies to the granulite facies, where no calcium garnet occurs, rather than to the hornblende amphibolite facies where it would be expected. The associated schists are of little value in making a subfacies assignment for these rocks, because the assemblage sillimanite-garnet locally characterizes the pelitic rocks through the upper part of the almandine amphibolite facies and into the granulite facies.

Two chemical analyses of hornblende-augite specimens from group *B* are plotted on an AC diagram (Fig. 5). These specimens show such

high CaO content and high MgO:FeO ratio (Table 14) that they are certainly not isochemical with the mafic bands of the gneiss, and direct comparison is not possible.

Port Deposit Granodiorite

The metamorphic minerals in the granodiorite in this area indicate metamorphism after emplacement. This is most clearly indicated by the widely distributed euhedral garnet porphyroblasts. Along with chlorite, epidote, and plagioclase (An_{15}-An_{32}), these indicate metamorphism at the lowest level within the almandine amphibolite facies. This assignment is in accord with Turner's recent suggestion that the boundary between the greenschist and almandine amphibolite facies be drawn at An_{10}, rather than at the difficult-to-determine boundary between hornblende and actinolite (Fyfe, Turner, and Verhoogen, 1958, p. 218). In the main outcrop area of the granodiorite in Cecil County, Maryland, the mineralogy of inclusions and wall rocks suggests greenschist-facies metamorphism (Hershey, 1937; Marshall, 1937).

Gabbro

The Bringhurst stock appears to be almost entirely the product of igneous processes and therefore is not classified under the metamorphic-facies scheme. However, the igneous gabbro facies (Eskola, 1939) represents physical conditions similar to those which gave rise to the surrounding granulites.

The Iron Hill stocks apparently originated by igneous processes. The widespread uralitization which converted much of the original pyroxene to actinolite and actinolitic hornblende appears to have resulted from post-crystalline thermal or hydrothermal metamorphism.

The formation of the uralite and the change in the extremely calcic plagioclase to give epidote plus a less calcic plagioclase probably do not represent metamorphic equilibrium but rather show incipient development of low-grade metamorphic conditions where subaluminous amphibole is a stable phase in an aluminous rock. The alteration of the peridotite portions of the Iron Hill rocks to serpentine was probably contemporaneous with the uralitization and represents the same low-grade metamorphic conditions as does the uralitization (Bowen and Tuttle, 1949). These two processes probably resulted from the emplacement of the surrounding Port Deposit granodiorite.

Summary

The metamorphic zones present are disposed roughly concentrically about the granulitic Arden granite, showing decreasing intensity of metamorphism from the center outward. The juxtaposition of facies and subfacies is generally in good agreement with Turner's recent revision of the metamorphic-facies scheme (Fyfe, Turner, and Verhoogen, 1958). Two points, however, might be noted. First, between the granulites and the upper almandine amphibolite facies materials there are no rocks which correspond to the newly erected hornblende granulite subfacies. Although hornblende-augite-hypersthene rocks are abundant and widespread, they give clear textural evidence of diaphthoretic origin and cannot be considered an equilibrium assemblage. Second, the amphibolites of groups *B* and *D* which do contain pyroxene (augite) and hornblende in apparent equilibrium occur in an outer zone separated from the granulite by several miles of hornblende-plagioclase amphibolites.

If these rocks are the equivalent of the former almandine-diopside-hornblende subfacies, as they appear to be, then that subfacies is not, in this area at least, clearly transitional between the almandine amphibolite and granulite facies.

Structure

Structural Setting of the Wilmington Complex

The Piedmont crystalline belt in this region is 40 miles wide. In the western portion the foliation strikes approximately east-west. In the central portion the foliation turns northward and strikes approximately N. 55° E. The over-all structure is poorly known, and what is known is subject to controversy. Geologic maps (Bascom and Stose, 1932; Stose and Stose, 1944; Swartz, 1948, p. 1503) reveal a series of broad anticlines and synclines which are defined largely on a stratigraphic basis. At the cores of the anticlines the Baltimore gneiss is exposed; the Glenarm series is exposed in the synclines. Mafic rocks (*i.e.*, bodies mapped as gabbro) are abundant in the anticlines, where they are associated with the Baltimore gneiss. Such bodies are much less abundant in the synclines. The southernmost syncline lies in the Wissahickon formation adjacent to the Wilmington complex (Ch'ih, 1950, p. 934). The complex

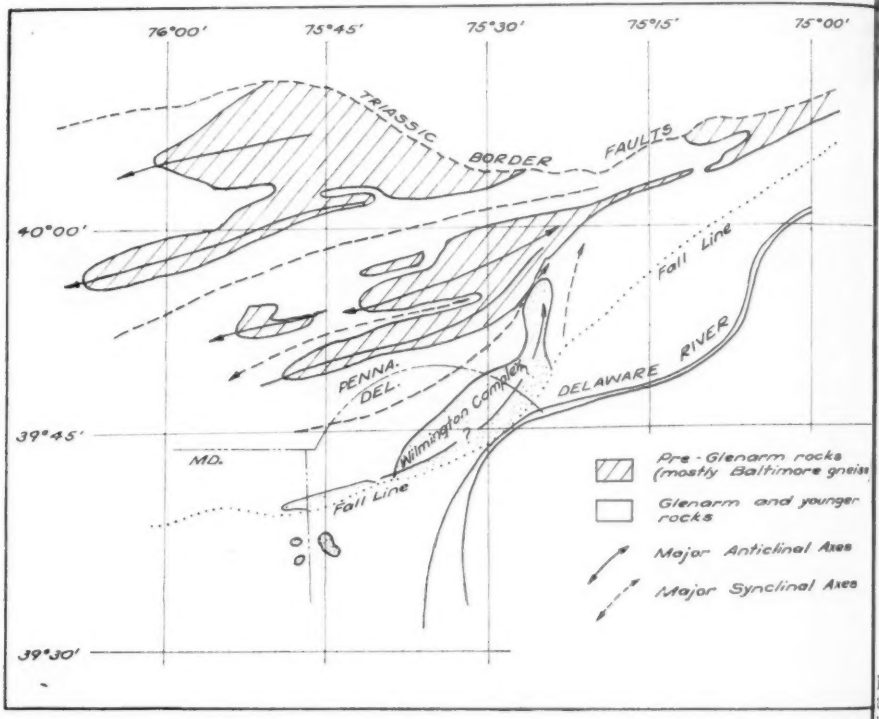

FIGURE 6.—STRUCTURAL FEATURES OF NORTHERN PIEDMONT

lies in what appears to be the next anticlinal belt to the south, which plunges beneath the Wissahickon formation at its northeastern end. Its southwestern end is obscured by the Port Deposit granodiorite and the overlap of Coastal Plain sediments (Fig. 6).

Foliation and Lineation

In the banded gneiss and the amphibolite of the complex two separate planar features can be detected. These appear to be fundamentally different and warrant separate description and consideration.

BANDING (S_1): In the banded gneiss S_1 is defined by the irregular alternation of the relatively thick mafic and felsic bands described previously. This banding is displayed at every large exposure, although the mafic bands may be represented only by fragments 1–2 feet long.

It is also present in the amphibolite and is defined by noticeable variations in the amount of hornblende or in the grain size of adjacent layers. Layers range from 4 inches to 6 feet thick but are most commonly about 1 foot thick.

FOLIATION (S_2): In both phases of the banded gneiss and in the amphibolite, S_2 is a distinct gneissic banding produced by the concentration of the component minerals in layers approximately one-quarter of an inch thick. Streaks or layers of lenticular quartz grains are also oriented in this plane. The foliation is more prominent in the amphibolite, where contrast between the dark amphibole and white plagioclase is great, than in the banded gneiss, where the pale pyroxene and the gray or blue plagioclase offer little contrast. However, close examination shows that S_2 in the gneiss is well formed and that the individual layers may be traced for several feet. Within the mafic bands of the gneiss irregularities of foliation are rare or absent, and the foliation is, in most places, parallel to the borders of the bands. The foliation in the coarser-grained felsic material

however, does not in all cases exhibit perfect parallelism with the contacts of the adjacent mafic bands. It commonly shows irregularities and swirls and in places may appear absent.

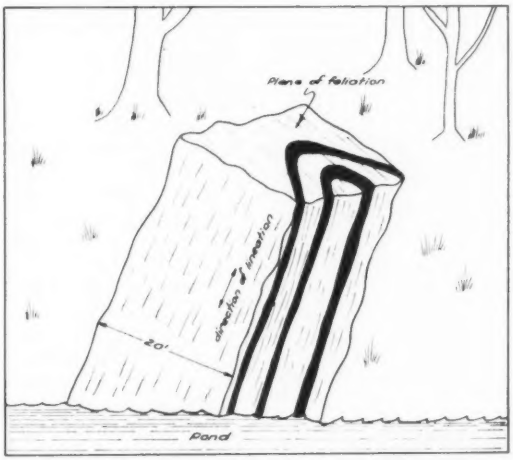

FIGURE 7.—OUTCROP SKETCH OF S_1 FOLD IN AMPHIBOLITE
Color contrast in adjacent layers has been exaggerated.

In most exposures S_1 and S_2 display complete parallelism. However, marked discordances of the banding and foliation are present in at least three large exposures, and in many small exposures slight discordances and ambiguous relationships occur. In one place (G113) a mafic band about 6 inches thick is drawn out in boudins where the foliation cuts across the crest of an S_1 anticline. At another place (D4) a band which makes an angle of about 75° with the foliation has been deformed in the plane of S_2, producing jagged edges, top and bottom, for several feet. The angle between S_1 and S_2 is rarely great, and discordances of a few degrees may have been overlooked in many instances.

In the amphibolite a marked discordance of S_1 and S_2 can be seen at a small quarry behind the New Castle County Workhouse, where S_1 lies in a steeply plunging fold with an exposed amplitude of about 20 feet. The foliation is parallel to the axial plane of the fold, cutting across S_1 sharply at the crest and more gently at the limbs. A mineral lineation parallels the fold axis, but it is weak because the rock has nearly granoblastic texture (Fig. 7).

Megascopically detectable linear features in the complex vary with the composition.

MINERAL LINEATION: In hornblende-rich rocks, lineations are distinct where hornblende grains are large and elongate and obscure in fine-grained, granoblastic specimens. In the banded gneiss, mineral lineation is commonly obscure because the principal components have equant habit. In some places, however, streaks of small pyroxene or quartz grains do define a mappable lineation, but more commonly apparent linear elements prove to be traces of planar features.

BOUDINAGE: In at least two outcrops which show the crests of folds in the banded gneiss exposed in a plane approximately perpendicular to the fold axis, the mafic bands are drawn out into boudins at or near the crest of the fold. The boudins are parallel to the fold axes and define a *b* lineation. The mineral lineation at these places is also in the *b* direction.

JOINTING: The rocks of the complex display one or more joint sets at every exposure, but the joints are not commonly so well formed or closely spaced as in the nearby Wissahickon schist. Distinct cross joints are the most abundant and widely distributed expression of jointing in the complex. In the amphibolite they are perpendicular, or nearly so, to the mineral lineation defined by the parallel hornblende grains. In the banded gneiss the cross joints are most common in the thicker mafic bands (6 inches thick or more), where they are oriented approximately perpendicular to the long dimension and to the mineral lineation,

where detectable. These run the full thickness of the mafic bands but rarely extend into the surrounding lighter material. They are present not only in the continuous bands but also in the larger remnants and boudins.

Relationship of Structural Elements

Whatever the origin of S_1, its present attitude is a result of deformation (folding). Portions of S_1 folds are commonly exposed and, more rarely, entire small folds can be seen. The foliation, S_2, is generally parallel to S_1, the banding, suggesting development in flexure slip folds. Where the foliation cuts across the crests of folds or shows discordance with S_1 on limbs, it may have originated by either or both of the following: the development of a structure similar to the fracture cleavage of the Wissahickon formation (S_3 of Cloos and Hietanen, 1941; Ch'ih, 1950) or flow of the felsic material during syntectonic metasomatism and mobilization. In the Wissahickon formation the fracture cleavage is apparent as small crinkles of the S_2 plane, but no such features have been detected in the complex. Investigation of fracture cleavage in massive rocks such as these would require petrofabric analysis, which was not undertaken for this study. That the felsic material has undergone some plastic flow is established by such features as intrusion into openings in the mafic bands, irregular swirls, small ptygmatic folds, and, rarely, of small areas where the felsic material appears massive.

Where exposures are such that the relationship between the linear elements and the folds in which they occur can be seen, the lineations are parallel to the fold axes, and it seems reasonable to assume that this is generally the case. The lineations are therefore considered *b* lineations (Knopf and Ingerson, 1938, p. 57). In intensely deformed rocks such as the Arden granite, *a* lineations may be present in the plane of shearing, but observations are inconclusive.

The structural features of the complex are shown on Plate 1. Attitudes of the foliation of the schist near the contacts are shown for comparison.

The foliation in the complex is generally parallel to its contacts with the schist and with the foliation in the schist. In the central and northern portions the foliation is vertical or dips steeply northwest. This trend is continued to the northeast along the strike in the Wissahickon formation (Ch'ih, 1950, Pl. 1, Pl. 6, fig. 91). In the southwestern portion of the comple the dips are to the south or southeast. Th trend is also continued along the strike in th granodiorite and metavolcanic rocks of Ce County, Maryland (Hershey, 1937, Pl. 2 Marshall, 1937, Pl. 35). The change fro southerly dips to northerly dips coincide approximately with a change in trend of the regional strike from east-west to about N.55°E

At the northern end of the complex there are places where foliation is not parallel to the contact but rather cuts across it sharply (Pl. 1). In these places, however, the foliation parallels that in the adjacent schist.

Within the banded gneiss there is a small area where the foliation shows relatively low dips and strikes which are in part discordant with the regional trend. These appear to form a low dome, at the center of which lies the Bringhurst gabbro stock. This structure is tentatively interpreted as a dome resulting from the emplacement of the stock.

Another area in which the foliation is markedly discordant with the regional trend occurs immediately north of the Delaware-Pennsylvania boundary. No explanation of this anomaly is offered here, but it should be pointed out that areas of similar trend have been mapped nearby (Ch'ih, 1950, Pl. 1).

The lineations and fold axes are usually horizontal or plunge slightly northeast. This is in agreement with petrofabric studies made in the Wissahickon schist in the region (Cloos and Hietanen, 1941, p. 157; Ch'ih, 1950, p. 936) which show an average northward plunge of 16°.

The lineation has a consistently steep northeasterly plunge in the amphibolite around Brandywine Springs, Delaware. The regional strike shifts here from east to northeast, and this area has been interpreted as part of a major regional structure marked by steep-axis folding (E. H. Watson, personal communication). This would presumably involve the flexure of the axial planes of earlier-formed folds, and the steep *b* lineations formed in the folds are to be considered distinct from the major *b* lineation of the region. Such a phenomenon has been described in the Poundridge area, New York (Scotford, 1956), and given the perhaps unfortunate term "axial-plane folding."

The schist which is interbedded with the amphibolite in the area of Rockland is directly in line along the strike with a larger area of schist which crops out in the vicinity of Brandy

ine Springs (Pl. 1). The more southerly area pelitic material is interbedded with, or included by, a few small sheetlike basic bodies, but the nature of the exposures is such that it has not been possible to determine whether the pelitic mass lies above or within the major portion of the amphibolite. However, along the southern contact, exposures of the two rock types are separated by less than 100 feet. In this instance the amphibolite dips below the schist at 40° (Pl. 1).

In summary, the complex exhibits such close conformity of structure with the Wissahickon formation as to indicate that they were deformed together. Furthermore, the general parallelism of the contact between the schist and the complex with the regional structures and with exposed stratigraphic contacts in the region appears to reflect a stratigraphic relationship: the complex generally underlies the Wissahickon formation, although minor interbedding occurs. Such a relationship is also suggested by the occurrence of the complex in what appears to be an anticlinal area bordered on the northwest by a synclinal belt. The changes in the outcrop pattern of these flexures as they are followed along the strike (to the northeast) are what would be expected by their gentle northeasterly plunge; the complex becomes narrower and ultimately disappears below (?) the schist.

Origin and History

Introduction

In this discussion the term complex refers mainly to the banded gneiss and the amphibolite contiguous with it. The Arden granite and the gabbro stocks are clearly of later origin and will be treated separately. The amphibolite bodies within the Wissahickon formation will also be treated separately.

Discussion of the origin of the complex must deal with the dark bands of the banded gneiss and with the amphibolite. The felsic portion of the gneiss which intrudes, and appears to replace, the mafic bands gives evidence of having arisen by metasomatism and intrusion during metamorphism. It will be dealt with more fully in the section on history.

Origin

Three characteristics of the complex which are premetamorphic in origin and presumably date from the original formation of the body appear to provide a basis for determining its origin. These are: (1) the persistent banding, S_1; (2) the over-all shape and structural relationship to the Wissahickon formation; and (3) the chemical characteristics of the rocks.

The most likely possibilities are that these rocks originally consisted of (1) a stratiform gabbroic body; (2) a series of sedimentary rocks; (3) a series of basaltic lavas or tuffs; or (4) a plutonic gabbro body. Each of these possibilities may be considered in turn.

(1) Lopoliths of the Skaargard or Stillwater type, which have resulted from differentiation, in place, of gabbroic magmas, are rich in FeO near the upper contact (Wager and Deer, 1939) and grow progressively richer in CaO and MgO downward (Hess, 1941). The exposed contact of the Wilmington complex with the Wissahickon formation appears to be its upper contact, but whether upper or lower it is neither sufficiently iron-rich nor calcium rich to be considered a result of *in situ* magmatic differentiation. Moreover, the progressive variation in composition with respect to distance from the contact which occurs in lopoliths has not been detected here.

It may be, however, that the body originally displayed these gradational chemical characteristics but that they were obscured or destroyed by pre-Glenarm erosion or later by metamorphism and the metasomatism which accompanied it.

(2) The complex may have had a sedimentary origin as a series of impure limestones or limestones and graywackes. Such an origin would adequately explain the persistent S_1, the shape of the body, and its apparent stratigraphic relationships. The rocks are, however, too poor in Al_2O_3 for graywackes on the one hand and too iron-rich and carbonate-free for limestones on the other. It might be argued that complex metasomatism during metamorphism produced marked changes in composition and that all CO_2 was lost at a time when the temperature was high and the confining pressure low. This possibility is remote in light of the fact that the Cockeysville marble at nearby outcrops is composed almost entirely of calcite and dolomite, demonstrating that its reconstitution was accomplished at pressures high enough to retain most, or all, CO_2.

A serious objection to a theory of sedimentary origin is that no sedimentary rocks of comparable composition occupy this stratigraphic position anywhere in the region.

(3) A volcanic origin of the complex, tuffs or flows, would provide an adequate explanation for the position and shape of the body and for the S_1 as bedding. The composition of the mafic bands in the gneiss and of the amphibolites falls within the range of olivine basalts. Volcanic origin would also provide a reasonable explanation of the absence of ultrabasic rocks and of the presence of the apparently amygdaloidal texture of some of the amphibolites. In many places basic volcanic rocks occupy a stratigraphic position below, or in the lower part of, the Glenarm series and so admit of possible correlation with the Wilmington complex (Jonas, 1924; 1928; Jonas and Stose, 1938; 1939; Marshall, 1937).

(4) Plutonic origin as an intrusive gabbro would adequately explain the chemical composition, olivine gabbro. The S_1, by this hypothesis, would be a result of flow and perhaps multiple magmatic injections during crystallization. Banding is not uncommon in gabbros (Bowen, 1928, p. 170). The over-all shape and apparent stratigraphic position of the complex suggest that, if it originated as a gabbro, it was emplaced as a large sheetlike body more or less concordant with the Wissahickon formation.

The present fine grain of the mafic bands and the amphibolite and the absence of any relict texture attributable to a gabbro require extreme stress and attrition of an originally coarse-grained rock. Such a process is not impossible, considering the very high grade of metamorphism. However, the complete destruction of original texture throughout a large gabbro pluton is probably rare.

Of the four possibilities outlined, only sedimentary origin appears wholly inadequate. The following section on the history of the complex is concerned for the most part with the processes which affected the body after original emplacement or deposition, so for the present it is not neccessary to make a choice from among the remaining possibilities.

History

Introduction.—Any account of the history of so complex and highly metamorphosed a body as this must be conjectural. As more work is done in the area or as concepts of petrogenesis advance, modifications will be necessary.

Banded gneiss and amphibolite.—In late Precambrian or early Glenarm time a series of basaltic lavas or tuffs was deposited in this area upon the Baltimore gneiss basement or in the lower part of the Glenarm series and subsequently covered by the Wissahickon sch and the formations above it. Alternativel sometime after the deposition of the Glena series a gabbroic magma was intruded eith hypabyssally, to produce a stratiform sheet, at depth, producing a typical gabbro pluto Ultimately deep burial took place, accompani by regional metamorphism. Recent age dete minations of the Wissahickon formation sugge that this may have had its culmination duri the Appalachian Revolution (Watson, 195 p. 2–5). The area now occupied by the compl lay very close to the orogenic axis (Hess, 195 p. 395) and underwent extreme metamorphism.

Locally, the highest-grade metamorphism occurs in the area occupied by the banded gneiss, where the granulite-facies level was reached. Roughly concentric areas of progressively lower metamorphic grade are developed outward from the banded gneiss. Recrystallization was accompanied by a migmatization which produced the felsic bands in the gneiss.

The migmatization process appears to have been metasomatic, differing from granitization principally in the potassium-poor nature of the material added. Probably less refractory feldspar-rich layers of the original mafic rock were selectively replaced by solutions or emanations consisting largely of Na_2O and SiO_2. This was accompanied by the removal of FeO, MgO, and CaO. Evidence for this is present in those slides of the felsic material where small groups of rounded pyroxene grains have identical orientation, indicating that they are partially replaced "xenocrysts" of originally more mafic material. All slides of felsic bands show pyroxene crystals embayed and partially replaced by quartz, and in some cases by plagioclase; these are apparently earlier or later stages of the same process. Throughout the felsic material, quartz grains are large and lobate, indicating origin during a different stage or cycle of crystallization from the associated granoblastic pyroxenes.

During the process the felsic material, in many places, became sufficiently mobile to appear truly igneous and invaded or engulfed and partially digested, the remaining mafic bands. Much of the mafic material, however, not only remained solid during the process but appears to have undergone little change other than recrystallization. The persistence over considerable distances of these sheetlike mafic bands indicates that S_1, the original banding or bedding, was a major controlling factor in the migmatization.

The chemical changes necessary to convert

afic band to a felsic one can be expressed more less quantitatively by using the method of e standard rock cell proposed by Barth 952). This method adopts the premise that etasomatism is a constant-volume process and

ABLE 15.—CATIONS ADDED TO AND SUBTRACTED FROM A MAFIC BAND (D4a) TO FORM A FELSIC BAND (D4b)

Added	Subtracted
2.7 ions of Na	8.7 ions of Ca
0.1 ions of Ti	13.5 ions of Mg
13.8 ions of Si	4.7 ions of Fe^{+2}
	1.6 ions of Al
16.6 ions added	28.5 ions subtracted
er 160 oxygen ions representing 58.3 valences	per 160 oxygen ions representing 58.6 valences

ompares equal volumes of material (*i.e.*, portions containing equal amounts of oxygen). Table 15 shows the cations which must be dded and removed from a standard cell of the afic material to produce an equal volume of he felsic material.

Mafic) $K_{0.2}Na_{2.5}Ca_{13.1}Mg_{15.9}Fe^{+2}_{7.0}Fe^{+3}_{1.3}Al_{16.7}Ti_{0.2}Si_{47.8}O_{160}$

Felsic) $K_{0.2}Na_{5.2}Ca_{4.4}Mg_{2.4}Fe^{+2}_{2.3}Fe^{+3}_{1.3}Al_{15.1}Ti_{0.3}Si_{61.8}O_{160}$

The figures given in Table 15 must be used ith caution for at least two reasons. The first s that the replacement may not have been on n equal-volume basis. In fact, it appears that considerable dilation has occurred. Secondly, s unlikely that the material replaced was dentical with the mafic bands which remain. Chemically, the felsic material differs from ublished analyses of trondhjemite only in that he felsic bands have somewhat less potassium Table 14).

That this process took place at extremely igh temperature is indicated by the fact that granulite-facies assemblages are restricted to he felsic material and to the mafic bands ssociated with it. To the southwest, where igmatization is absent, the metamorphic rade does not exceed amphibolite facies.

Arden granite.—The Arden granite, which is ocated near the center of the granulite-facies rea, has characteristics which indicate that it as formed by the action of a potassium-rich magma or granitizing agent upon the banded gneiss and which also indicate that it intrudes the gneiss.

This apparent dilemma is resolved by considering the body a parautochthonous granite (Read, 1951). By this hypothesis, the granite was formed by sodium and potassium metasomatism (granitization) of pre-existing rocks, in this case the banded gneiss, in a zone of intense regional metamorphism. During the process it became sufficiently mobilized to move (upward) with respect to the surrounding rock, thereby acquiring intrusive relationships to it, or at least destroying the highly attenuated, completely gradational, contacts which accompany *in situ* granitization.

The relationship between the potassium-poor granite, or trondhjemite, of the felsic bands and the potassic granite at Arden appears to be this: during regional metamorphism normal granitic igneous activity was in progress at some depth below the level now exposed. (Whether this was migmatization and granitization at the edges of granite magma or whether it was the migmatization and granitization which produced the granite magma is a problem which is unresolved by the rocks exposed here.)

At distances from the center of activity so great that no migmatization took place, the basic rocks reached only the amphibolite-facies level. At somewhat lesser distances the material introduced was potassium-poor (trondhjemitic) and produced the granulite-facies banded gneiss. At greater depth, where true granitization was in progress, potassium was abundant, as were the heat and stress necessary to mobilize a portion of the rock so affected. As a result of this mobilization the potassic Arden granite became emplaced in the migmatitic gneiss which lay above it. Observations have been made by others supporting the view that sodium-rich, potassium-poor emanations or magmas travel farther from the centers of activity than normal granitic materials. In the Lake Manapouri terrain in New Zealand Turner (1937, p. 236–237) reported that ". . . . the granitic rocks carrying 10 per cent or more of potash-feldspar tend to be concentrated toward the centre of the injection-complex while trondhjemites are more frequent toward the eastern and western margins". This phenomenon is explained in part by the observation that the sodium ion is more mobile than the larger potassium ion (Lapadu-Hargues, 1945, p. 225). The emplacement of the Arden granite appears to have been contemporaneous with the regional metamorphism.

Gabbro.—The massive gabbro stock at Bringhurst has not been deformed or recrystallized and is therefore thought to have been emplaced well after the peak of regional metamorphism. It is unlikely that it was intruded during an entirely new cycle, because it has undergone minor diaphthoresis and metasomatism identical with that in the enclosing gneiss.

The Iron Hill stocks near the southwestern end of the complex are probably contemporaneous with the Bringhurst stock, but the facts that these bodies are widely separated and that the contact relationships of the Iron Hill bodies are unknown leave the problem largely unsolved.

Port Deposit granodiorite.—The Port Deposit granodiorite, which intrudes the amphibolite at the western end of the complex and probably surrounds the Iron Hill stocks, exhibits structural and mineralogical characters which have resulted from progressive metamorphism (*i.e.* strong foliation, garnet porphyroblasts). End-stage solutions from the granodiorite have invaded the amphibolite *lit-par-lit* or as cross-cutting pegmatites and locally caused marked diaphthoresis. These apparently anomalous relationships indicate that the granodiorite must have been emplaced early enough in the period of regional metamorphism to have become seriously deformed and also that some igneous activity associated with the granodiorite took place after the complex had reached its present metamorphic grade. A large compound batholithic body such as this cannot have developed rapidly, and indeed, igneous or metasomatic activity in the granodiorite probably extended over the entire tectonic cycle.

It seems likely that heat or hydrothermal solutions associated with the introduction of the granodiorite and pegmatite into the amphibolite provided catalysis for the uralization of the Iron Hill gabbro and that the granodiorite may be in part older than and in part younger than the bodies at Iron Hill.

Diaphthoresis and metasomatism.—The final events recorded in the rocks of the Wilmington complex are the retrograde metamorphism which affected most of the granulite-facies rocks and the adjacent amphibolite by replacing the pyroxenes with hornblende and the minor potassium metasomatism by which small quantities of biotite were introduced into most rocks of the complex. The fact that this biotite replaces the later hornblende is evidence that this was the last phase to crystallize. It might be argued that the diaphthoresis and the potassium metasomatism could have taken place at any time between the original metamorphism and the final denudation of the complex, but there is strong evidence that these processes actually took place well before the end of the main metamorphism. Mineralogically, both the hornblende and the biotite are varieties which crystallize at higher metamorphic grades. The hornblende is of an aluminous green or brown-green type which occurs in all high-grade rocks of the region, and the biotite is a yellowish-brown to reddish-brown, high-grade variety, rather than a green or black green one, which is indicative of the lower-grade processes (Wyckoff, 1952, p. 39).

Amphibolites of the Wissahickon formation.—These bodies range so widely in size and mineralogical character, and the area in the Wissahickon formation examined during this investigation represents so small a part of the whole formation, that no definite conclusions can be drawn about the origin of these bodies. The possible origins are, again, sedimentary, volcanic, and intrusive, and it seems not at all unlikely that, in the Wissahickon formation, all three origins are represented. An additional possibility, hybrid origin, appears to be represented in one specimen (F70). This scapolite- and calcite-bearing rock is skarnlike and suggests a basaltic magma contaminated by limestone or a magmatically altered limestone.

Possible Correlations

Although it has not been necessary to choose among the four postulated origins of the mafic material of the complex in order to discuss its structural and petrologic evolution, it would be of value to the history of this portion of the Piedmont to know the nature of the original material. On the admittedly weak ground that there is no evidence to the contrary, and also that there is some positive evidence, the author is inclined to suggest that the original material was volcanic.

McKinstry (1949, p. 877–879) wrote, after working in the schists adjacent to this complex,

"... Also partly or wholly surrounded by Wissahickon are bodies of basic rock that have been mapped as gabbro and resemble some of the mafic material in the Baltimore gneiss. It is conceivable that some, though by no means all, of this material is metamorphosed basic lava, as no other equivalent of either the Catoctin greenstone or the basic volcanics of Maryland have been recognized east of the Susquehanna".

As mentioned previously, a group of meta-volcanic rocks with amygdaloidal texture ha

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en described from Cecil County, Maryland. the nearest point these are about 10 miles ng the strike from the amphibolites of this mplex. Marshall (1937, p. 205, 212) estimates eir total thickness as exceeding 1 mile and signs them to a stratigraphic position below e Wissahickon formation, although they are irely surrounded by the Port Deposit anodiorite. If the Wilmington complex were rt of the same volcanic series, the necessity having this great thickness of volcanic cks in Cecil County stand alone, isolated by osion, would disappear. The original volcanic tures are better preserved in Cecil County an they are in this complex as a result of the wer grade of metamorphism (greenschist as ainst amphibolite and granulite facies in the ilmington area).

A question might also be raised as to whether olcanic rocks in Cecil County, Maryland, and the Wilmington area are related to the lavas South and Catoctin mountains, which lie out 100 miles to the west at the other edge this portion of the Piedmont. Such long-nge correlation would have to be extremely ntative, but it is noteworthy that the atoctin-South Mountain volcanic rocks are, part, normal calcalkaline metabasalts and cupy a stratigraphic position similar to that cupied by the basic rocks in this area. Cloos Cloos and Hietanen, 1941, p. 102) has signed the Catoctin-South Mountain rocks to e Cambrian because they overlie the Pre-ambrian Baltimore gneiss and are in turn erlain by fossiliferous lower Cambrian rocks hich are a portion of, or correlative with, the lenarm series. The volcanic rocks were pre-iously considered Precambrian (Jonas and ose, 1938). Structurally the Catoctin-South ountain rocks display complete conformity ith the overlying rocks, a characteristic which shared by the Wilmington complex.

A relationship worthy both of mention and of rther investigation is this: some of the major ticlines of this portion of the Piedmont con-in, or are bordered by, linear belts of basic cks, appearing on the maps as discontinuous odies of gabbro surrounded by either Balti-ore gneiss or Wissahickon schist. These bodies e several hundred yards to 1 mile wide and to 5 miles long. These basic rocks may be utcrops of an originally continuous sheet lying tween the Baltimore gneiss and the Glenarm ries. Such a body might be correlative with th the Catoctin series and the Wilmington mplex. Its discontinuous outcrop pattern ight be the result of post-Catoctin erosion or the thrust faulting which is evidenced by the fact that in many places the Wissahickon formation lies directly upon the Baltimore gneiss.

The suggested relationships are highly speculative, but they do indicate that a volcanic origin for the Wilmington complex would simplify some regional relationships. The stratigraphic and structural relationships of other "gabbros" of the Piedmont must be clarified by more detailed field and laboratory work before the regional problems can be solved.

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LEGEND

IGNEOUS ROCKS

- Alluvium
- Pillow lavas and tuffs
- gd — Granodiorite and diorite
- amp — Amphibolite
- py — Pyroxenite
- Nodular serpentinite
- gn — Fine-grained serpentinite (greenstone) with and without microscopic actinolite

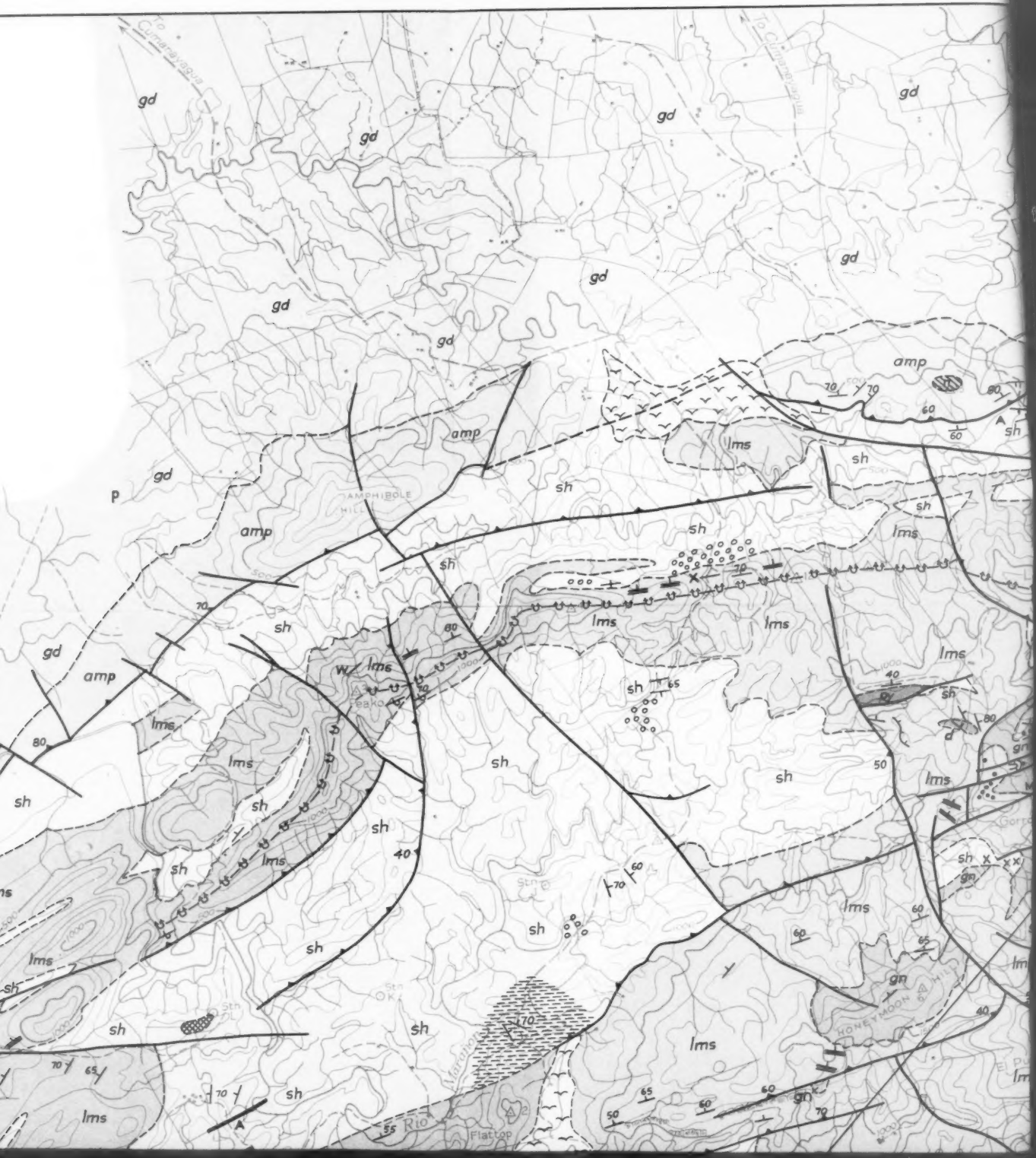
SCHISTS

- sh — Undifferentiated carbonaceous "chloritic mica" schists
- ooo — Garnet-muscovite schist
- Black argillite
- Manganiferous schist
- Graphitic garnetiferous schist
- x x x — Epidote-quartz-mica schist

CARBONATE ROCKS

- Dolomite horizon
- d — Dolomite
- lms — Undifferentiated limestone

- Geological boundary (generally faulted and inferred)
- Strike and dip — 55
- Strike, dip presumed overturned
- Strike, vertical dip
- Strike, dip unknown
- Fault, dip as indicated — 60
- Fault, assumed
- Axis, overturned antiform
- Gossan
- Magnetite-chlorite schist (locality) — M
- Brucite limestone (locality) — W, X, Y, Z
- Actinolite vein — A
- Road
- Track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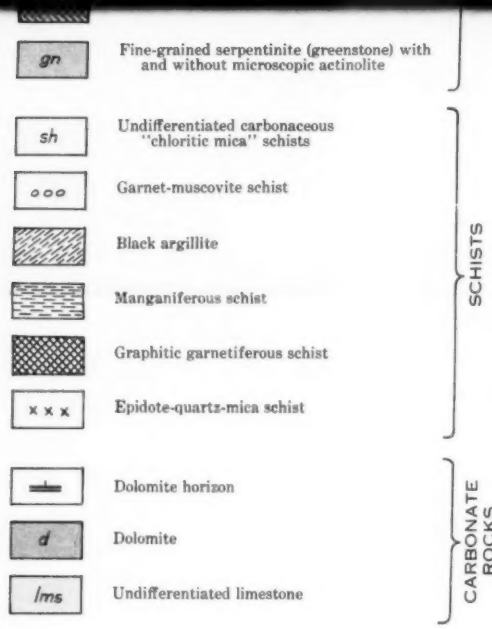

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Strike, vertical dip
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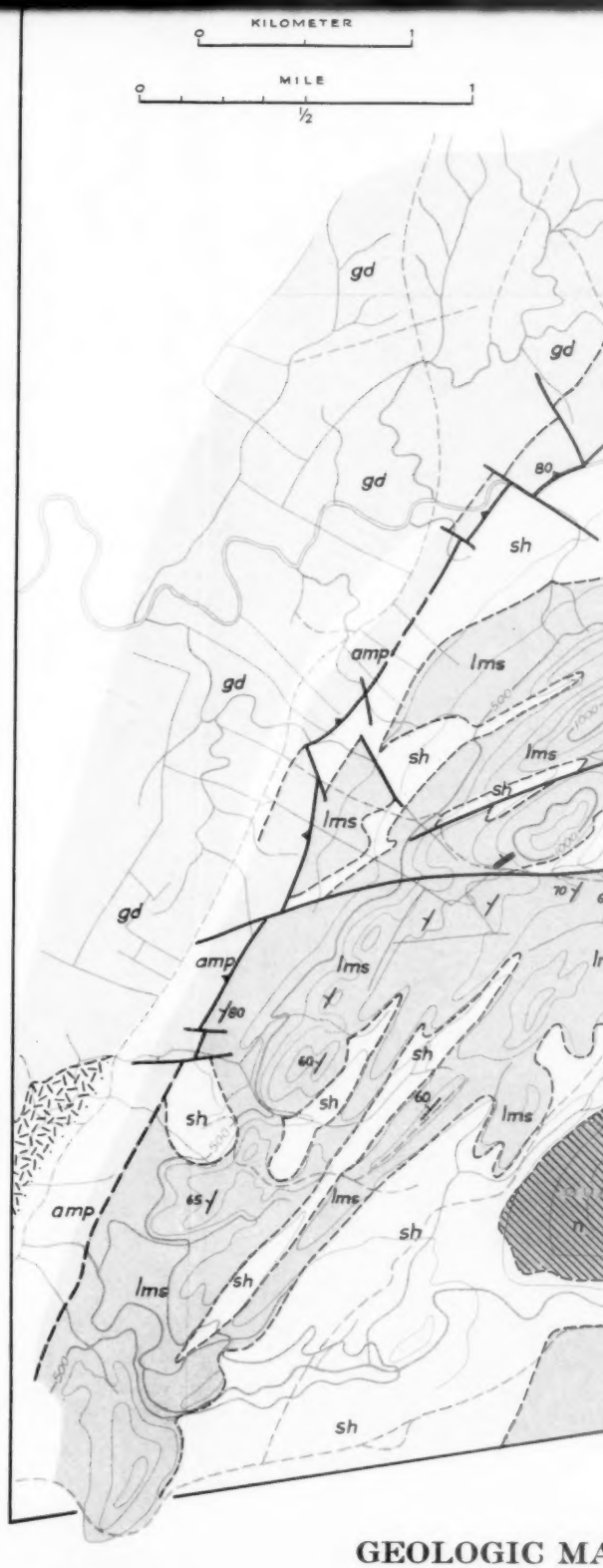
Fault, assumed
Axis, overturned antiform
Gossan
Magnetite-chlorite schist (locality)
Brucite limestone (locality)
Actinolite vein
Road
Track
Fence
Building
Drainage
Triangulation point
Important plane-table station
Contours

NOTES

Dip and strike symbols for bedding are centered in circular sample cells of approximately 1000 square feet area, and where angle of dip is given, represent the means of 5 to 10 replicated readings rounded off to the nearest 5°. Dip figures for faults, with the exception of the amphibolite contact, were obtained photogrammetrically.

Geology by Hill 1951, 1952-1953. Area between triangulation points plane-tabled by Hill and Lewko 1952-1953. Photogrammetric extension 1954-1955. Precision deteriorates westward, especially west of P-Q. Topographic names in English are provisional only. Revised and corrected 1959.



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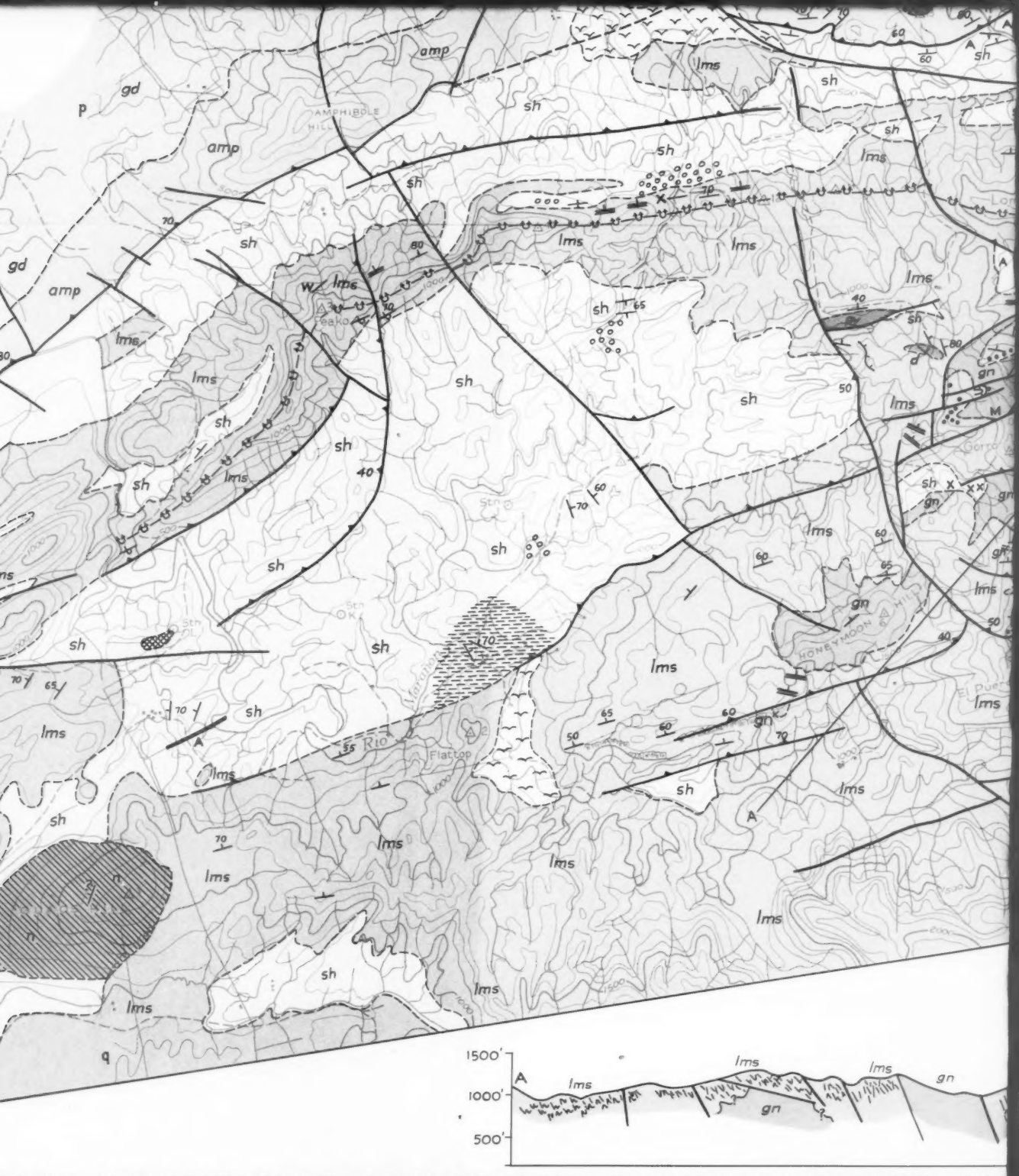
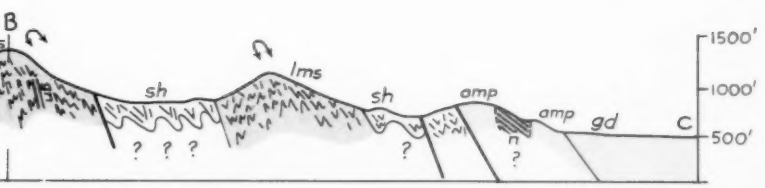


E MAP AND SECTION OF THE NORTHWEST TRINIDAD MOUNTAINS, LAS

NORTHWEST TRINIDAD MOUNTAINS, LAS VILL



LLAS PROVINCE, CUBA

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GEOLOGY AND STRUCTURE OF THE NORTHWEST TRINIDAD MOUNTAINS, LAS VILLAS PROVINCE, CUBA

BY PATRICK ARTHUR HILL

ABSTRACT

The major rock types of the Trinidad Mountains of central Cuba are carbonate rocks and schists of presumable Mesozoic age. Both are carbonaceous, yet devoid of fossils. The carbonate rocks comprise limestones (about 90 per cent), brucite limestone, and dolomites.

The brucite limestone contains chloritic mica (sheridanite). The dolomites, despite lenticulation, might serve as marker horizons. Two varieties are noted, one with crystals of black euhedral gypsum, the other without them.

Most widespread of the schists is a carbonaceous chloritic mica schist which alternates with quartz-garnet-mica schist or less commonly grades into epidote or talc schists.

Serpentinites are of two types: a nodular type derived from peridotites, and a fine-grained type (greenstone) derived from microgabbro. The greenstone is older and has been affected by movement of two periods; one (preserpentinization) formed actinolite, the other (postserpentinization) locally developed magnetite-chlorite schist. Other chloritic derivatives include a goethite-chlorite schist and an albite-lawsonite-chlorite schist. The predominant chlorite is clinochlore.

All rocks are isoclinally folded. Cleavage or jointing that might serve to establish overturning is absent. Strike faults are prominent. Transverse faults, because of the lack of traceable units, are difficult to recognize.

The mountains appear to form a gigantic anticlinorium. A continuous band of amphibolite averaging 1200 feet in thickness separates the carbonate rocks and schists of the mountains from diorite and granodiorite to the north. Although thrusting may exist within the mountains, the large northerly directed thrust formerly postulated for the northern boundary is disproved.

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Introduction

The Trinidad Mountains of central Cuba (Fig. 1) lie almost wholly within La Villas Province and include the sierras of Trinidad, San Juan, and Sancti Spiritus. They rise to a maximum height of 3792 feet (Pico Potrerillo) and consist of isoclinally folded limestones and schists, which in the area of Plate 1 dip steeply north and northwest. The rocks lack fossils and are of unknown age. However they are older than the dioritic intrusions to the north and the Cretaceous sediments to the west (Thiadens, 1937).

In some areas, serpentinites associated with the limestones are accompanied by pyrite. The largest pyrite deposit is at Minas Carlota (Hill, 1958b).

Field work was conducted by the author alone during July and August 1951 and with the assistance of Mr. Edward Lewko of Uranium City, Saskatchewan, from December to February 1952–1953.

Laboratory work comprised examination of thin sections and mineral separations, together with X-ray and spectrographic analyses. A Debye-Scherrer-type camera was used for the X-ray work. A 3-meter Baird grating-type spectrograph with 300-volt, 6-ampere D.C. arc was used for spectrographic analyses.

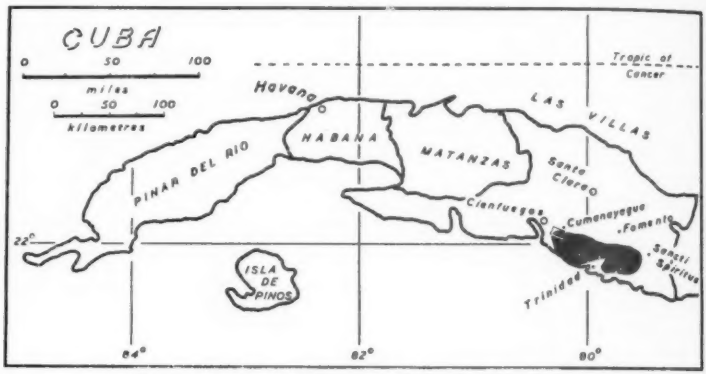

FIGURE 1.—LOCATION OF THE TRINIDAD MOUNTAINS

Acknowledgments

The study was sponsored by the Freeport Sulphur Company and the Department of Geology, Columbia University; both were generous with financial assistance and equipment.

Thanks are due the Alonso family of Cuba their hospitality and also to Mr. Rich Colligan, the Diaz de Villegas Company, the late George Sicklick.

Technical help was rendered by Mr. W Sutcliffe of the University of Nottingh England, and Mr. John Stewart of the M Branch, Ottawa. Professor W. D. Gill m available laboratory facilities at Trinity Col Dublin. Mr. F. W. Dunning of the Geolog Survey of Great Britain painstakingly identi or confirmed many minerals in thin sections separations.

A special debt is owed Prof. Charles Behre, Jr., for advice and encouragem Profs. Paul F. Kerr, Ralph Holmes, and Sid Paige read parts of the manuscript. Dr. D de Vletter and Dr. Charles W. Hatten supp editorial criticism.

Previous Regional Investigations

Thiadens (1937) noted that "A rather l amount of literature concerning Cuban geo exists. Notwithstanding, the geologic knowl of the island in general and also of the sout Santa Clara province is very poor." In the geological map of Cuba, de Castro (1 represents the schists of the Trini Mountains as Silurian-to-Carboniferous s mentary rocks, bounded on the north granite, serpentine, and basalt.

Spencer (1896, p. 71) notes that althoug Castro assigns the formations of the Trini Mountains to the Paleozoic

> "...there appears to be no certainty as to t age. They are composed of a semicrystalline

stone of fine texture and blue colour, resembling externally some Ordovician rocks, so also do the associated black slates . . . but in their internal structure the Trinidad mountains constitute the oldest rocks of central Cuba."

Hayes, Vaughan, and Spencer (1901) regard the schists as basement rocks. Wright (1916, p. 8) mentions that the town of Trinidad was founded in 1513 because gold was discovered in the Rio Arimao. Calvache (1925) reports that large pyrite deposits were found in the sixteenth century south of Cumanayagua; presumably these are the Minas Carlota (Pl. 1). Allende (1928) describes these as ellipsoidal ore bodies conformable in "Precambrian" schists.

Thiadens (1937) summarized a survey undertaken 4 years previously by Dutch students. This paper contains an excellent descriptive bibliography and reconnaissance map. Describing the rocks of the Trinidad Mountains, Thiadens writes:

"Geologic history begins with the genesis of those rocks which furnished by metamorphism the schists and marbles of the Schist Formation. The age of these rocks cannot be stated with certainty, as fossils if formerly present, entirely disappeared in consequence of the metamorphism. By comparison of these rocks with others found elsewhere on Cuba and on surrounding islands, it may be deduced that the age is very probably Lower Cretaceous and Upper Jurassic. This age, however, has not been proven and the rocks may be very well older. They are at least of pre-Middle Cretaceous age."

Other writers also speak of a Schist formation. In this study the "schists" are divided into carbonate rocks and schists. The term schist formation is not used hereafter.

Geomorphology

The serrated sky line (Pl. 2) suggests that the Trinidad Mountains consist predominantly of limestone. Field work within and without the area of Plate 1 confirms this.

The limestone forms high ground. The prominent northerly dip gives hills a distinctive silhouette with a gentle northern slope and a steep southern one. Karst features are common, including dry valleys, caves, sink holes, and, on a lesser scale, solution cavities. Many valleys end abruptly in steep walls, and many are crossed at right angles by rock buttresses. Caribbean karst lands are described by Ahmann, Krömmelbein, and Lötschert (1956) and Sweeting (1958).

Soil is red, especially where the adjacent limestone is schistose and rich in hematite. Soil creep is rapid; thus the best exposures of limestone are usually on high ground.

The original drainage of the northwest Trinidad Mountains was probably to the north and west, governed by the dome structure of the mountains. The next stage, superimposition upon underlying bedrock, is indicated by wind gaps along the Gorro-Chivo ridge which are aligned with streams draining off El Puerco.

The third and present stage of drainage was the rapid development of subsequent streams along strike faults or easily eroded belts of schists. The Navarro River (Pl. 1) which has captured many consequent streams is an example.

River terraces are rare. They are locally developed at Charco Azul. Higher-level terraces that cut directly across bedrock are more common and are found at 1400 feet at Station *B* (Pl. 1), at 1250 and 1600 feet on La Barona (Pl. 2, fig. 2), and at various levels between 1300 and 1500 feet on Honeymoon Hill.

Upraised and tilted peneplains of unknown date are indicated by concordant summit levels (Pl. 2, figs. 1, 3).

Carbonate Rocks

General Statement

The carbonate rocks are dynamically metamorphosed, devoid of fossils, and are believed to be pre-Middle Cretaceous (Thiadens, 1937). They are crystalline, foliated, or schistose. All are graphitic; all are steeply and isoclinally folded. They include limestones (comprising about 90 per cent of the carbonate rocks), dolomite, and brucite limestone.

An attempt was made to locate fold axes. Because of the absence of fossils, lithology was used for correlation. Each natural, unbroken outcrop along the Cumanayagua-Carlota road was studied (Pl. 4). The following limestone varieties were found:

(1) Light-blue densely compressed limestone
(2) Medium-blue massive-weathering platy and foliated limestone
(3) Medium-blue massive-weathering foliated limestone
(4) Limestone weathering light blue, foliated, and containing black calcite crystals up to an inch in diameter
(5) Flaggy limestone with bluish-white patina
(6) Similar to 5 but with schist bands
(7) Tightly folded limestone ranging from white to blue; may be equivalent to (1)
(8) Schistose limestone with whitish patina

(9) Dolomite, buff to light gray, with kneaded and sculptured appearance. One band with small black gypsum crystals

(10) Brucite limestone resembling weathered lamprophyre in color and appearance

Although these lithostratigraphic subdivisions confirmed isoclinal folding, only one fold axis (the northern axis through Chivo) was discovered. Attempts to discover other axes were unsuccessful because (1) only the dolomite and brucite limestone, which together form less than 10 per cent of the carbonate rocks, are distinctive enough to be recognized in any outcrop; (2) the limestones proper are subdivided by subtle shades of color that are not sharp enough for stratigraphic distinction; (3) isoclinal folding is so intense that different rock types alternate abruptly; (4) rock flowage along strike is extreme, so that some units thin to disappearance; (5) cleavage and jointing are virtually absent and cannot be used to determine overturning.

The most reliable exposures are on hillsides where runoff is rapid and rock faces dry. Where runoff is slow or continuous, water seeping through the overlying mantle coats outcrops with a buff-pink patina, or it may deposit layers of gray tufa several inches thick. Consequently exposures in many stream beds, particularly where the stream is small or intermittent, are useless for mapping, because bedrock and loose rock fragments are cemented together and coated with thick sheets of travertine.

Limestones Proper

The outcrops of rocks of lithologies 1–8 are, in hand specimen, composed of equigranular relatively massive limestone and laminated limestone. These, perhaps, are equivalent to the crystalline and the quartz-bearing limestones of Thiadens (1937). Each is divisible into subtypes, between which every gradation exists as a result of (1) change in proportion of minerals, (2) change in grain size, and (3) flowage, injection, and mingling of units. The equigranular and laminated limestones are subdivided macroscopically as follows:

Equigranular limestones

E1 Black dense limestone with "diabasic" texture; grain size less than 1 mm

E2 Speckled limestone intermediate between E1 and E3

E3 Mottled blue-gray and white limestone with augen of blue limestone averaging ½ to inch in length, engulfed by white calcite

Laminated limestones

L1 Silvery micaceous and schistose limesto containing white calcite in many ca stained yellow with iron, hematite cu pseudomorphic after pyrite, and sider veinlets commonly less than 1 inch thick

L2 Gray schistose limestone of chloritic a pearance; graphite content stains finge visible pyrite

L3 Red schistose limestone, slightly hard than L1 or L2, with abundant red hemat powder along laminae, micaceous lus present but not prominent

In general the equigranular limestones yi massive outcrops. Flaggy outcrops are develop where equigranular and laminated limeston alternate.

Both the equigranular and laminated types a veined by calcite. Siderite is rare. In hand specime quartz veinlets are inconspicuous, but some ou crops are cut by quartz veins up to 2½ feet thick

Microscopically the difference between equ granular and laminated limestone as seen about 50 slides is one of degree only; each grad imperceptibly into the other. The subtyp appear to depend upon the varying proportio of the constituents.

EQUIGRANULAR COMPONENT: Sections show granoblastic mosaic consisting of:

(1) Calcite, comprising at least 85 per cent of t rock and showing twin lamellae, usually two s for each crystal, commonly bent. The lamell generally incline at 30°–50° to any foliation prese Much of the calcite is slightly biaxial and clou gray to yellow owing to a high proportion of grap itic or ferruginous inclusions. In places the inclusio are concentrated into rectangular or six-sid patches indicative of a pyrite or epidote crys partly or wholly replaced.

(2) Quartz, averaging 5 per cent and invariabl present either as rounded, elongated, or subangul grains or as irregular patches or fractured veinl in places crenulated and indicating introduced recrystallized quartz. Under low magnification gas or liquid inclusions are visible, but extre high-power magnification shows glassy pink to clusions of indeterminate composition.

(3) Muscovite, averaging 3 per cent, as cle laths of varying length scattered through sections. No alignment is recognizable.

(4) Hematite, translucent and blood red, pse domorphic after rectangular pyrite.

(5) Pyrite, surrounded by hematite.

(6) Magnetite, found in separations altho not distinguished in microsection.

(7) Graphite which, although suspected inclusions in calcite, is recognizable only in se rations.

(8) Apatite, occasional, commonly broken and shredded.

(9) Epidote or clinozoisite in columnar aggregates associated with iron-rich bands. Polarization colors are low-order dark grays; the optical sign is biaxial negative; the interference figure is poor; and two cleavages are in some cases present at 70°–90° angles with one another. The possibility of plagioclase cannot be ruled out.

(10) Sphene and chlorite as occasional constituents of the iron-rich bands.

The different varieties of equigranular limestone result from variations in size of the calcite and variations in amount and distribution of inclusions. The mottled subtype is an injection phenomenon with light calcite surrounding and veining blue-gray calcite. Vein calcite is everywhere white, suggesting that in recrystallization and injection it has been cleared of inclusions.

LAMINATED COMPONENT: The textures of sections examined range from gneissic-granoblastic to intensely crumpled, swirled, and microfaulted. The sections contain:

(1) Calcite, composing at least 85 per cent of the material seen in the slides. Individual grains are more elongated than in the equigranular variety; some grain boundaries are "fused" together and difficult to recognize. Twin lamellae are wavy and multiple.

(2) Quartz, poikiloblastic in calcite or as angular veinlets or lenticles along definite folia. No detrital quartz was seen, and this may be diagnostic of the laminated variety. There is more quartz in subtype L3, and it is arranged in lenticles.

(3) Hematite and goethite compose most of the abundant red and brown powdery or mossy aggregates scattered throughout the calcite. Replacement of pyrite is especially noticeable in subtype L3.

(4) Muscovite, everywhere present in amounts of about 5 per cent. It may be arranged sporadically or in parallel or subparallel folia. In many places it is corrugated and associated with dark graphitic powder, and it may occur in swirl aggregates with quartz and calcite (as a pseudomorph after garnet?).

(5) Chlorite, of two types. One is colorless, twinned, with lamellae at right angles to schistosity, and commonly occurs in the same folia as quartz. Polarization colors are dark gray (slightly yellow) to black. No anomalous blues are seen. The second variety is pale green, nonpleochroic, scaly, with cleavage at high angles to schistosity. Its extinction is parallel; it is length slow and uniaxially negative with no rings or colors; its polarization color is low and blotchy. The laminated varieties of limestone contain more chlorite-serpentine than the equigranular varieties.

Accessory minerals are apparently not so numerous as in the equigranular varieties, but this may not be diagnostic.

Dolomite

Dolomite outcrops are buff to light gray and appear kneaded and sculptured. Their thickness ranges from 2 to 20 feet and averages 5 to 10 feet. Although lenticular dolomite was found only at Wilson Ridge, three-quarters of a mile west of Charco Azul, the Carlota variety, like many others, may be highly lenticular. The accompanying maps show only the actual outcrops; many of these undoubtedly connect under cover.

In hand specimen this rock is gray, fine-grained, and has a dense gritty texture. Veinlets of quartz less than 1 mm thick are common. Pyrite cubes occupy some of the graphite-chlorite folia in drill-core samples. Microscopically the rock consists of a sugary mosaic of anhedral calcite and euhedral to subhedral dolomite. Muscovite is scattered throughout the slides. Hematite occurs either as dust or pseudomorphic after cubes and possible intergrown pyritohedra of pyrite.

Perfect euhedral crystals of black gypsum up to 5 mm long are scattered throughout one band of dolomite 6 feet thick on a road bend north of Charco Azul. This is at least locally a useful marker bed.

Unfortunately the small gypsum crystals pull out when slides are made. However, thin sections show that the groundmass of the gypsiferous rock consists of cloudy fine-grained patches of calcite, rare muscovite, and abundant graphite surrounded by euhedral to subhedral dolomite with interstitial quartz and sporadic thin muscovite laths, all with a slight graphitic content. No hematite was seen.

Approximately 45 gms of the rock was dissolved in warm dilute HCl. The residue on filtration amounted to 5.8 per cent and, as confirmed by F. W. Dunning, consisted of black euhedral gypsum 1–5 mm long, containing clouds of dark inclusions, biaxially positive with $2V = 60°$ and n_z approximately 1.530, and chloritic mica with n_z approximately 1.57 and containing dark inclusions. No anhydrite was detected in either of these two subtypes.

Brucite Limestone

An interesting brucitic limestone is exposed near limestone-schist contacts. Because of soil cover, outcrops could be traced only a few feet along the strike. This unit appears to be 1–4 feet thick. The rock is brownish black, highly graphitic and dusty, and laminated and

speckled with laths (up to 10 mm long) of lustrous black brucite. This, too, is a useful horizon marker.

Microscopic sections (Pl. 3, fig. 1) reveal a murky fine-grained aggregate consisting of brownish-black graphitic powder, poorly defined calcite granules, minute calcite veinlets, and very rare muscovite in small broken laths. The graphitic powder is arranged in parallel or subparallel lines and may occur in fuzzy rods at an angle of 25° with each other. The detail in the calcite is not clear, even under high power.

Irregularly inset in the calcite-graphite groundmass are elongated subhedral porphyroblasts of brucite, uniaxial and optically positive, with zoned extinction and anomalous blue-gray interference colors. Smaller, usually clearer, crystals show good longitudinal cleavage with a suggestion of another cleavage at 75° to this. Graphitic inclusions are prominent parallel to crystal outlines, especially to the longer dimensions. Generally the inclusions are a continuation of the graphitic lines of the groundmass. The brucite appears to have grown after the foliation of the rock developed.

To supplement the thin sections the following studies were conducted. Samples averaging 6 gms were dissolved in acid. The percentages of filtrate obtained from localities w, x, y, and z (Pl. 1) were respectively 17.8, 8.78, 28.8, and 15.4. The residues revealed (1) abundant black flakes closely resembling graphite but identified by F. W. Dunning as chloritic mica (biaxial, optically positive, 2V very small, n_z 1.57) containing hosts of minute dark inclusions, some of which may be carbonaceous, others ferruginous; (2) brownish fragments of hydrated iron oxide, possibly goethite, together with some finely divided micaceous material; and (3) minor amounts of magnetite, hematite replacing pyrite, muscovite, irregular grains of quartz, and possible undissolved brucite.

As a check on optical identification the chloritic mica was X-rayed. Two runs were made, the first, copper radiation with a nickel filter, for identification; the second, cobalt radiation and an iron filter, for photographic reproduction purposes. The chlorite was found (Tables 1; 2) to be sheridanite (McMurchy, 1934). The sheridanite was also spectrographically analyzed. Carbon, suspected from the optical analyses, was confirmed as a minor constitutent.

Halloysite

Although the limestones of the Trinidad Mountains are subjected to tropical weathering, aluminous residual products are absent. Heavy rainfall and high relief, which result in heavy slope wash, are responsible. Pockets of bauxite may exist, but only where very unusual circumstances favor their preservation.

The only aluminous product discovered was a chalky white powder which encrusts an outcrop of limestone along the road leading to El Capitañ (Pl. 1). Microscopic examination indicates it is halloysite, extremely fine-grained and with a refractive index slightly greater than balsam.

The material was X-rayed. Three runs were made, the first, iron radiation for identification; the second, iron radiation and manganese filter, for visual intensity determination; and the third, cobalt radiation and an iron filter, for photographic reproduction (Table 3).

The clarity of the X-ray pattern is good. Although several lines appear that are not recorded by at least 6 authors, the extra lines agree closely with those of the Mines Branch, Ottawa, Canada.

PLATE 2.—PANORAMAS

FIGURE 1.—Serrated limestone topography. Panorama looking west from Salto del Hanabanilla about 10 miles east of Minas Carlota. Note northerly slope of the dissected upraised peneplain which forms the sky line (*cf.* Fig. 3).

FIGURE 2.—View from dolomite knob three-quarters of a mile west of Charco Azul, looking eastward toward Chivo (Peak 9) and La Barona (Peak 16). Two high-level terraces with elevations are shown.

FIGURE 3.—Panorama from Peak 5 looking southwest toward Guaos Hill (Peak 1) and the coastal plain. On the left, an upraised peneplain surface slopes gently toward the right. Banding is visible in the Guaos Hill serpentinite.

FIGURE 4.—A panorama, complementary to Figure 3 of this plate (building *X* is common to both photographs) from the saddle south of Gorro looking northwest across open, hummocky, schist country toward Peako (Peak 3).

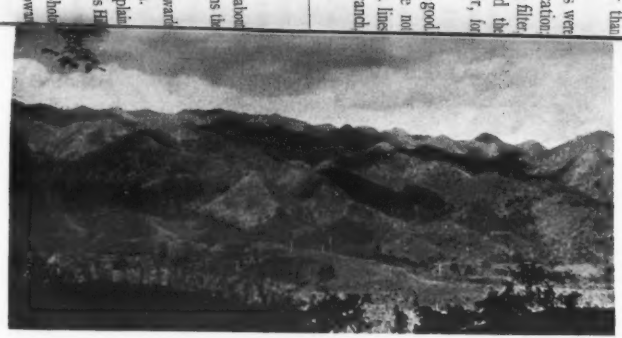

FIGURE 1

FIGURE 2

FIGURE 3

FIGURE 4

PANORAMAS

FIGURE 3

FIGURE 6

FIGURE 9

FIGURE 2

FIGURE 5

FIGURE 8

FIGURE 1

FIGURE 4

FIGURE 7

The material was also analyzed spectrographically. Semiquantitative comparisons were made against standard calibration powders. The results are listed in Table 4.

The possible significance of the halloysite cannot be considered here. No alunite or endellite were detected. The halloysite appears to have been deposited by ground water issuing between red clay and limestone. Its purity may indicate selective leaching or selective precipitation.

SCHISTS

General Statement

Schists of sedimentary origin occur within the map area (Pl. 1) and to the east and south. As with the limestones they lack fossils, are tightly isoclinally folded, and usually possess steep northerly dips.

Schist topography is smooth and gently rolling (Pl. 2, fig. 4). Soil is gray or silvery from the carbonaceous or micaceous constituents. Thiadens (1937) mentions quartz-mica schists with accessory garnet. At least seven varieties of schists exist.

Carbonaceous Chloritic Mica Schist

These schists are soft, soil the fingers, and weather to form gray soil. Fresh specimens suitable for thin sectioning are difficult to find Mineral separations show that the chloritic mica of these schists resembles that of the limestones in containing minute dark inclusions, some of which may be carbonaceous, others ferruginous. This is the commonest schist. It may grade into, or with isoclinal folding alternate with, quartz-mica-garnet schist.

Quartz-Mica-Garnet Schist

Where carbonaceous and associated with chloritic-mica schist, this schist is easily weathered. By itself, it may form high ground; it occurs high up the north and south sides of Peak 4 (Pl. 1). The rock is corrugated and lustrous with silvery muscovite. The garnets vary in amount and size, are red, and give the rock a spotted appearance.

Two varieties exist: variety 1 occurs throughout the area; variety 2 is restricted to the Marañon River north of Flattop (Pl. 1).

Variety 1 consists of lenticular sheared and sliced quartz, albite, and muscovite laths with

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PLATE 3.—PHOTOMICROGRAPHS

FIGURE 1.—(×30) Brucite limestone. Brucite laths in a groundmass of graphitic dust and calcite granules. Note clear strips in brucite parallel to groundmass. Locality 1', Plate 1

FIGURE 2.—(×9) Garnet in graphitic quartzite. The euhedral garnets follow rock lamination. Cross-bedding is suggested by merging 1' at top left

FIGURE 3.—(×12) Magnetite skeleton in felted chlorite (clinochlore), cut by shear. Locality: Wilson Ridge, half a mile north of Gorro

FIGURE 4.—(×40) Goethite-chlorite schist. Biotite with cleavage at high angle to lamination, replaced by chlorite; quartz with inclusions (Q); apatite showing cross fracture. Locality: Charco Azul

FIGURE 5.—(×12) Albite-lawsonite-chlorite schist. Albite (Ab) and lawsonite in a chloritic groundmass. The sphenes (black) of the groundmass pass through the lawsonite crystal. Locality: gossan approximately 10 miles east of Charco Azul

FIGURE 6.—(×12) Serpentinite; antigorite with magnetite. Note relict of pyroxene outlined by finer-grained matted antigorite and outlines of tubby lath-shaped mineral ophitically contained in the upper right of relict of pyroxene. Locality: Guaos Hill

FIGURE 7.—(×24) Olivine-pyroxene rock. Cross-polarized light. Large first-generation (?) pyroxene embayed and surrounded by olivine, pyroxene, and amphibole with lines of included magnetite and green spinel in the large pyroxene. Locality: Terminal Hill

FIGURE 8.—(×18) Amphibolite showing incipient alteration. Granulation in hornblende and plagioclase (An_{50-60}) with sheaves of anthophyllite developed along zones of microshearing. Locality 4, Cumanayagua-Minas Carlota road (Pl. 4)

FIGURE 9.—(×50) Amphibolite showing more advanced alteration than Figure 8. Sheared porphyroblasts of clinozoisite are set in a groundmass of nephrite, sericite, and altered plagioclase (lower left). Locality 5, Cumanayagua-Minas Carlota road (Pl. 4)

occasional porphyroblasts of garnet. Rock deformation is indicated by curvature of the albite and swirl structure in the muscovite.

TABLE 1.—X-RAY POWDER LINES OF SHERIDANITE
(Cobalt radiation: iron filter)

I	*d*(*kX Å*)
6	13.79*
10	7.07
5	4.74
4	4.16
1	3.70
9	3.53
2	3.35 (graphite)
4	2.84
5	2.59
8	2.55
6	2.44
4	2.38
5	2.25
1	2.06
9	2.00
3	1.88
2	1.81
½	1.74
½	1.71
1	1.66
4	1.56
8	1.53
2	1.50
½	1.46
1	1.41
3	1.39
3	1.32
2	1.29

* Obtainable only on long film (camera diameter 114.3 mm)

The fabric suggests some postcrystallization deformation.

The albite is generally corroded and subhedral, in some cases euhedral. Subhedral crystals contain quartz and other constituents of the groundmass. Euhedral crystals form straight or curved blades and may contain dusty inclusions (graphite?) arranged in bands parallel to their length. Simple Baveno twinning and rare lamellar twinning are present. The garnet has developed in and includes the groundmass, the rotation and distortion of which it faithfully reproduces. Hematite forms metallic granules and spangles. No magnetite was identified.

Variety 2 consists of garnet porphyroblasts in a groundmass of lenticular calcite. In th groundmass, muscovite laths follow lamin and wrap around garnets; quartz shows wa extinction, angular outlines, and is older th

TABLE 2.—STANDARD X-RAY POWDER LINES FOR SHERIDANITE
(From McMurchy, 1934, p. 423–424; maximum intensity—10; camera radius 57.3 mm. CuK_{α} 1.537)

I	*d*
7	13.678
8	7.040
2	5.250
9	4.680
2	3.922
10	3.509
1	3.129
7	2.828
½	2.712
7	2.578
5	2.542
6	2.430
3	2.370
4	2.247
½	2.215
1	2.064
7	2.021
6	1.998
4	1.881
4	1.825
3	1.731
3	1.706
2	1.659
10	1.562
10	1.534
2	1.502
2	1.460
3	1.417
10	1.390
2	1.350
4	1.319
4	1.283
½	1.251
7	1.218
½	1.191
3	1.181
4	1.131
4	1.094
5	1.040
5	1.030
3	1.013
3	1.001
4	.987

he calcite. Minor apatite in columnar aggregates, scattered sphene granules, and possible rounded dolomite follow the rock lamination.

Inset in the groundmass are: porphyroblasts of large euhedral garnet containing quartz

TABLE 3.—X-RAY POWDER LINES OF HALLOYSITE

(Cobalt radiation: iron filter)

I	*d*(*kX* Å)
10	7.33
9	4.44
8	4.12
2	3.72
1	3.60
2	2.55
2	2.34
1	1.67
5	1.48
1	1.28
1	1.23

TABLE 4.—SEMIQUANTITATIVE SPECTROGRAPHIC ANALYSIS OF HALLOYSITE

(Analyses by Mines Branch, Ottawa, Canada)

Locality	Elements—per cent													
	Fe	Si	Al	Mg	Mn	Pb	Cr	Sn	Ni	V	Ca	Cu	Zn	Ti
White chalky powder encrusting limestone on roadside near El Capitañ, Minas Carlota	0.1	25	15	0.02	less than 0.005	0.02	less than 0.002	less than 0.01	0.05	less than 0.01	less than 0.01	less than 0.005	less than 0.05	less than 0.002

clusions reproducing the swirl structure of e groundmass; large augen of pale-green non-ochroic diopside (2V about 50°) altered ong the margins to calcite and to yellowish-een chloritic glass; and scattered laths of ochroic (colorless to pale-blue) glaucophane hich appear to replace shredded ends of opside crystals and to be associated with uscovite.

Black Argillite

By increase in carbon content, the carbonaceous chloritic-mica schist grades into black rgillite. Good exposures are at Charco Azul d southwest of La Barona (Pl. 1). North of attop the argillite is dark brown and yields lamina sections for manganese. All argillites are ows wavy tremely weathered; most are soft, crumbly, lder than d damp and could not be sectioned.

Graphitic Quartzite with Garnet

Loose slabs of tough laminated quartzite, averaging 5 by 5 by 3 feet, lie apparently horizontally at plane-table station *L*. Weathered surfaces exhibit a dull-black ribbing of manganese oxide along the laminae which range in width from a fraction of a mm to 2 or 3 mm. Fresh surfaces are siliceous and show that the laminae are actually stratification planes with such features as cross-bedding.

Thin sections reveal predominant quartz, garnet, and manganese oxide. The quartz forms a mosaic of elongated grains, some of which are surrounded by a greenish film; others lack this film and show mortar structure. Small scattered rounded grains suggest that the original detrital shape of the quartz grains may be preserved.

The garnets are minute, yellow, and euhedral (Pl. 3, fig. 2) and contain inclusions. They are aligned along certain laminae and appear to have developed in an impure sandstone.

The manganese is black, in many cases colloform with brown translucent edges, and may be associated with carbonaceous material. It appears to be a late introduction, as it cross-cuts rock lamination and coats and replaces garnet.

Siliceous Graphitic Schist

These are silver-gray to grayish-black schists that soil the fingers. Unweathered specimens are rare and have a gritty feel. The rock has rough, slabby, jointing parallel to and cutting across the schistosity. Steep crosscutting joints are reported from similar (?) rocks in the Isle of Pines (Page and McAllister, 1944). Lineation is present along planes of schistosity and is caused by alignment of muscovite laths.

The rock is the only one in which pyrite weathers out instead of being replaced by iron oxides. Weathered outcrops are rusty, with yellow streaks and holes representing leached pyrite. The rock occurs in an area up to 1 mile

away from the graphitic quartzite into which it grades by increase in silica.

Thin sections show quartz (40–45 per cent), carbonaceous and ferruginous dust (30–35 per cent), muscovite (10–15 per cent), garnet, and pyrite pits (10 per cent). The texture is highly plicated, with slippage along the limbs of the plications. The garnets are small and colorless, generally decomposed, and, apart from a marginal dusty layer, do not affect the surrounding minerals. With higher magnification the dusty material proves to be brownish red.

Epidote-Quartz-Mica Schist

These schists are interbedded with argillites and serpentinite schists. The association and "purple" color—common to many tuffs elsewhere—may indicate a tuffaceous origin.

Under the microscope the rock consists of patchy and recrystallized quartz following schistosity; muscovite in wisps and laths; epidote in knots and augen, cloudy with dusty hematite and carbonaceous material and forming granular aggregates or crude six-sided porphyroblasts; hematite in red lenticles, the corners filled with yellowish-green isotropic glass; and light-green pennine (?) in veinlets with fibers at high angles to walls of veinlets. The pennine is crosscut by muscovite and appears to be older.

Age of the Carbonate Rocks and Schists

The limestones and schists of the Trinidad Mountains are of unknown age. Allende (1928) believed them to be Precambrian. Thiadens (1937) correlated them lithologically with schists in the Isle of Pines and in the lower San Andres formation of Pinar del Rio. As the upper San Andres was supposedly Upper Jurassic-Lower Cretaceous (Vermunt, 1937), he regarded them as pre-Upper Jurassic.

The lower San Andres formation (Cayento) has recently been dated as late Dogger (Krömmelbein, 1956). To assume that the rocks of the Trinidad Mountains are necessarily of the same age would be unwise. More than 50 per cent of the Cayento formation rocks are sandstones, which on metamorphism would form quartzites—common on the Isle of Pines (Rutten, 1934; Page and McAllister, 1944) but rare in the Trinidad Mountains.

Within the mountains the age relationship between limestone and schist is unknown. The limestone is depicted as older (Pl. 1), because limestone ridges as indicated by fold noses are anticlinal, and schist lies directly underneath younger amphibolite. However the ridges may be antiforms (Bailey and McCallien 1937), and the juxtaposition of schist and amphibolite may be fortuitous.

Faulting, isoclinal folding, and overburden prevent an immediate solution to the problems. For the present, the limestones and schists are regarded simply as "Mesozoic".

Igneous Rocks

General Statement

Igneous rocks of the Caribbean region are discussed by Schürmann (1935), Weyl (1950), Mitchell (1955), and Butterlin (1956). Analyses and descriptions of rocks from southern La Villas are given by Schürmann (1936) and Thiadens (1937).

The igneous rocks here discussed are immediately adjacent to, or within, the Trinidad Mountains. The diorite to the north is not described.

Although with one exception contact effects are lacking, the igneous rocks are considered younger than the enclosing limestone and schist. Nothing indicates that the sediments now represented by limestone and schist were deposited unconformably on an igneous basement.

The igneous rocks are divided into (1) fine-grained serpentinites and their schistose and chloritic derivatives, (2) nodular serpentinite and olivine rocks, and (3) amphibolites.

For the present, two groups of serpentinized rock are distinguished, based on field, megascopic, and microscopic characteristics. Each group may include rocks different in age, origin or degree of serpentinization. In particular the second group is broad and includes rocks which contain olivine or relicts of olivine.

All the igneous rocks contain much magnetite as dust or, less usually, as granules and stringers. Because of the fine particle size separation is difficult. Approximate estimates range from 7 per cent for the amphibolites to 25 per cent for veins in the Guaos Hill serpentinite.

Fine-Grained Serpentinites

Rocks of this category correspond to the serpentine schists of Thiadens (1937) and the greenstones of earlier reports. They stretch eastward from Honeymoon Hill through the main Carlota Valley and comprise green fine-grained felted rocks which originally may have been different but are now serpentinized and chloritized. Where massive these serp-

inites form high ground, but where schistose they form depressions. Contact metamorphism and jointing are absent.

Serpentinization of the fine-grained type may vary vertically, laterally, or in part according to the proximity of faults. Drill cores from the gossan area north of Gorro (Pl. 1) show that although outcrops appear completely serpentinized, the rock below may be only partially so and may contain unserpentinized actinolite. Accordingly, on the basis of surface information, serpentinization is described as incipient, partial, or complete.

INCIPIENT SERPENTINIZATION: Incipient serpentinization is found only at Honeymoon Hill where actinolitic serpentinite forms the core of the mountain. The contact on the north indicates a northerly dip of 30°; the southern contact is steeper, but the exact dip is unknown. The body appears to be a pluglike intrusion with an elliptical outcrop.

Macroscopically the serpentinite is a tough, dark-green rock with rusty phenocrysts. Polished sections are green and white with polygonal iron stains and are of igneous appearance with diabasic texture. Relicts of a larger mineral (plagioclase of pyroxene?) are apparent.

Thin sections confirm the relict structure and show rectangular areas composed of actinolite, zoisite, garnet, calcite, plagioclase, and rutile. The actinolite is pale green with brownish triangular cross sections. The mineral outlines rectangular areas as border laths but also may fill them. The border laths are commonly idiomorphic. The zoisite forms brownish aggregates. Individuals are too small for positive identification and may include some idocrase. Antigorite and chlorite are present in places as part of the zoisite aggregates. The garnets form euhedral crystals scattered through the zoisite areas. They may be arranged concentrically or clotted into polygons, the outlines of which may replace the cleavage of a former mineral, possibly pyroxene or amphibole. The calcite and plagioclase (An_{15-25}) occur in minor amounts and are usually associated with clinozoisite. Former lamellar twinning in plagioclase is suggested by lines of inclusions and clear areas in the brown zoisite areas. Rutile and possible perovskite are found in scattered murky-brown areas associated with iron staining.

The relationship between actinolite and serpentine is confused. Thin sections suggest that serpentinization followed actinolitization. However, (1) elsewhere, *i.e.*, in the faulted amphibolite, actinolite is the latest mineral and (2) movement appears to have been repeated along the east-west faults, so that actinolite may be of different ages.

PARTIAL SERPENTINIZATION: Increased serpentinization is shown by the rock immediately south of Gorro. The north contact is faulted; the fault plane dips 70° north. Other contacts are covered. On Peak 8 the rock is soft and light green, shows a glassy felted texture, and contains reddish-brown streaks.

Thin sections reveal that pennine and serpophite have increased in amount. Talc also has increased but is erratically distributed. Plagioclase (An_{10}) and sphene are associated with irregular patches of murky-brown rutile.

COMPLETE SERPENTINIZATION: Serpentinization is complete on Wilson Ridge and near Gossan 2 and is also indicated at Gossan 1 and 3 by loose float. The differences between these rocks are best considered by describing the two extremes—samples from Gossan 2 (one quarter of a mile north of El Puerco) and Wilson Ridge (half a mile west of Charco Azul).

Hand specimens of gossan 2 serpentinite resemble those from south of Gorro, but they contain more blue-green patches (averaging 5 mm in diameter) of structureless serpophite.

Microscopically the rock is a meshwork of fibrous antigorite, clear patches of serpophite, occasional lenticles and bursts of minutely felted talc needles, veinlets of magnetite dust, and scattered spangles of red hematite. Darker cloudy yellowish-green areas show finer-grain antigorite pseudomorphic after actinolite. The square outlines with truncated corners of these relict areas suggest that the actinolite was itself replacing an earlier mineral, probably a pyroxene.

At Wilson Ridge, development of fine-grained serpentinite, as indicated by surface outcrops, reaches its maximum. The rock is soft, apple green, and minutely felted with thin veinlets (maximum width, 1 mm) of magnetite dust. Picrolite, confirmed by X-ray analysis, is found at one locality associated with magnetite (Hill, 1958b, p. 969).

Thin sections reveal a uniform crisscrossed antigorite. Isolated relicts of former minerals are outlined by finer-grained, darker antigorite. Magnetite is ubiquitous as dust, stringers, and aggregates.

Schistose Serpentinites and Chloritic Schists

Schistose and chloritic derivatives exist (1) along sheared margins of large serpentinite bodies, *e.g.*, south of Gorro (Pl.1), (2) where

serpentinite is thin and sheared, and (3) where serpentinite grades into carbonaceous schist.

TALC-CHLORITE SCHISTS: These are partly equivalent to the chlorite schists of Thiadens (1937). They are soft, of greasy feel, with red ferruginous laminae, and contain minute matted fibers of tremolite. Drill cores reveal that they are commoner than surface outcrops would indicate. Some may have developed along horizons of sheared dolomite.

MAGNETITE-CHLORITE SCHIST: Greenish chlorite rocks with octahedra (up to 7 mm long) of black, shiny magnetite occur along the margins of many serpentinites. The largest outcrop is on Wilson Ridge. It may be traced laterally for 150 feet. Ferruginous laminae indicate schistosity dipping 40°–60° north; this is approximately the same dip as that of the surrounding rocks.

Thin sections show intermeshed tufts of pale pleochroic yellow to green chlorite, biaxially positive (2V = 5°–10°) with good cleavage, parallel extinction, and occasional anomalous polarization. X-ray powder lines for this chlorite are fuzzy but agree with those for clinochlore.

The chloritic groundmass contains large euhedral magnetites and magnetite skeletons (Pl. 3, fig. 3), occasional columnar apatites, rare idocrase (?), slightly bluish with extinction up to 10° and biaxially negative, and minute grains of broken epidote. Rectangular relict areas are discernible under crossed nicols. Minor antigorite fills the ferruginous laminae noted in hand specimens and may replace tremolite-actinolite.

GOETHITE-CHLORITE SCHIST: A waxy, yellowish rock with goethite pseudomorphous after striated pyrite cubes (up to 1 cm long), quartz lenticles, and reddish hematite laminae is exposed in Charco Azul (Pl. 4).

Thin sections show that the yellow color is derived from chlorite after biotite, films of serpentine (Pl. 3, fig. 4), and rare talc. The variety of chlorite could not be identified; X-ray powder lines are too fuzzy.

The chlorite infiltrates the biotite, which is pleochroic from colorless to yellow green. Cleavage in the biotite is at high angles to that in adjoining laminae; thus if oriented specimens could be collected, it might be possible to determine the relative direction of rock movement.

The quartz is lined with solid inclusions, is margined with apatite, and contains colloform goethite. It appears to have been introduced after the pyrite. Also present are colorless, idioblastic garnets, euhedral zircons, aggregates of dark-brown sphene and yellow rutile, and sporadic talc.

The goethite is translucent reddish brown and was identified by X-ray analysis. Ljunggren (1958, p. 23) describes schists where the chlorite formed probably "as a result of the decomposition of biotite by the action of sulphuric acid formed from the partial decomposition of pyrite".

The schist and the amphibolite (to the north) both have a high visible titanium content and much introduced quartz. However, the possible relationship between these two rock types cannot be settled on the basis of one small outcrop, for TiO_2 and SiO_2 are both mobile constituents.

ALBITE-LAWSONITE-CHLORITE SCHIST: Chlorite schists with porphyroblasts of albite and lawsonite occur approximately 10 miles east of Charco Azul. The schist is exposed about 200 feet south of a pyritic gossan, strikes in the same direction (east-west), and appears to dip steeply northwards. It is at least 8 feet thick. The rock is green with crystals of white hexagonal albite and gray rectangular lawsonite.

Chloritic tufts parallel or at high angles to the over-all schistosity (Pl. 3, fig. 5) constitute the groundmass. Relicts of amphibole are outlined by tufts at 120° angles. The chlorite is pale green, weakly pleochroic, and extinguishes obliquely at 2°–5°. Refractive indices are difficult to obtain; the chlorite was identified by X-ray analysis as clinochlore.

Irregular granules of sphene follow the crumpled rock laminae and run through both lawsonite and albite. Small crystals of apatite, needles of actinolite, and clean and minute garnets and bluish anatase (?) are scattered throughout the groundmass.

Albite porphyroblasts are colorless, euhedral and riddled with needles of actinolite. The porphyroblasts show simple and lamellar twinning and have 2V = 80° (approximately). The lawsonite is gray and usually rectangular. No actinolite is seen in the lawsonite, which appears to have formed before the albite. Incipient albite is possibly represented by patches of the groundmass that show uniform polarization.

Lawsonite-chlorite schists occur northeast of Santa Clara (Schürmann, 1936). The formula for this lawsonite is given as $2H_2O \cdot 2Si$... $Al_2O_3 \cdot (Ca, Mg, Fe, Na_2, K_2)O$. Harker (19... suggests that lawsonite is perhaps the ... step in the formation of zoisite or epid... from anorthite.

Nodular Serpentinites and Olivine Rocks

Members of this group contain olivine or serpentinized relicts of olivine. The group includes rocks different in age and perhaps in origin, including (1) the Guaos Hill serpentinite, (2) the amphibolite serpentinites, and (3) the olivine pyroxenite north of Wilson Ridge.

GUAOS HILL SERPENTINITE: At a distance, this serpentinite body (Pl. 1, Pl. 2, fig. 3) shows a layered structure dipping approximately 30° west. At close range this layering is not apparent, and only a rough sheeting, broken by joints at right angles, is visible. Nevertheless stereoscopic study of aerial photographs strongly suggests that the apparent dip is more than an illusion. Schürmann (1935, p. 345) and Vassal (1956) report layered gabbro-serpentinite elsewhere in Las Villas Province.

Weathered surfaces are nodular, appear somewhat waxy, and show variegated coloring. Freshly broken surfaces expose pinkish-yellow nodules (1 cm maximum length) set in a gray-green matrix. The rock is highly magnetic. With a lens, thin veinlets (1 cm maximum thickness) of dusty magnetite are visible.

Veins up to 18 inches thick of a smooth yellow serpentinite cut the nodular serpentinite. In these veins magnetite is conspicuous (up to 5 per cent) in wavy stringers. Many veins are concealed; limited exposures prevent accurate mapping.

Contact metamorphism of the surrounding rock appears absent, for although the actual contact is concealed, limestone and schist about 50 feet from the inferred contact show no metamorphism.

Thin sections show aggregates of serpophite and antigorite. The serpophite is yellow, occupies patches within the antigorite, and contains cores of magnetite. It appears to replace olivine.

Antigorite ($2V = 85°$) occurs predominantly as a mesh aggregate but also as veinlets. Former minerals are outlined by a finer-matted antigorite (Pl. 3, fig. 6). Many of the pseudomorphic areas show uniform and parallel extinction and may represent rhombic pyroxenes. Other nonmatted areas of antigorite show only traces of amphibole and pyroxene cleavage. Rare pleochroic tufts of yellowish-green bastite occur with the antigorite. Magnetite is present both in serpophite and as dust in veinlets and grids.

Yellow vein serpentinite proves in thin section to consist of fine-grained felted antigorite with scattered strains of magnetite dust. Bastite, with typical texture after rhombic pyroxene, is well developed. Large areas of magnetite are ophitic toward antigorite, and larger clots surround brown chromite. No picotite was detected.

AMPHIBOLITE SERPENTINITE: Amphibolite serpentinite is the name here given to partially or completely serpentinized igneous rocks cropping out along hilltops in areas of amphibolite. The original rocks appear to have been olivine pyroxenites. These have been serpentinized. The problem is which came first—serpentinization or amphibolitization?

Although contacts between the serpentinized pyroxenites and the surrounding amphibolite are hidden, the unaltered condition and the magmatic texture of the amphibolite strongly indicate that it is younger and engulfs serpentinized remnants of older rocks. Thin sections suggest that the green amphibole in these remnants is a late introduction.

Under the microscope the partially serpentinized pyroxenite consists of large pyroxenes embayed and surrounded by olivine, smaller pyroxenes (second generation?), and an amphibole.

The large pyroxene crystals (Pl. 3, fig. 7) are gray and nonpleochroic ($2V = 60°$). They contain plates and rods of magnetite and green spinel in planes parallel and at right angles to the longitudinal cleavage of the pyroxene.

The olivine, with low first-order polarization, forms rounded subhedral crystals and broad veinlike clusters. Wide twin bands parallel crystal length. The smaller pyroxenes form short subhedral crystals and show weak pleochroism from colorless to pale green. The amphibole is pale green and weakly pleochroic; it is length slow with a large 2V angle and has a maximum extinction of 15°–20°. It is anhedral and interstitial. Veinlets of yellow serpophite, granules of magnetite, and green spinel compose the remainder of the groundmass.

The more highly serpentinized rock is variegated yellow green with black veinlets. Thin sections show colloform serpophite and rippled magnetite infiltrating and surrounding mineral components. Fine, parallel-tufted antigorite occupies distinct areas; hematite has increased in amount.

OLIVINE PYROXENITE: Olivine pyroxenite crops out approximately 1 mile north of Gorro (Pl. 1). It is overlain by limestone which dips 40° north. Although the contact is partly covered, metamorphism of the limestone is

apparent—the only example of contact metamorphism discovered.

The pyroxenite is a tough, dark-green rock with salmon-colored ellipsoidal olivines (maximum dimension 2 by 5 mm). Weathered surfaces are brownish.

Microscopically the rock is hypidomorphic-granular with 35 per cent augite, 35 per cent hornblende, and 25 per cent olivine. Despite the lack of serpentinization, microsections resemble in many respects those from the serpentinized pyroxenites of the amphibolite areas.

The augite, 2V = 45°(?), is pale gray and contains included rods and plates of green spinel and magnetite. These appear to bisect the rectangular cleavage intersections and to make a 70° angle with many of the longitudinal cleavage traces. The olivine is subhedral and stained reddish yellow. Minor antigorite and chlorite are grouped together at augite interfacial angles. The hornblende is anhedral, shows schiller structure, and is pleochroic from pale yellow (α) to light green (γ). It appears to have been the last mineral to crystallize.

Above and grading into the pyroxenite is a garnet-pyroxene rock at least 4 feet thick. The garnets are euhedral, in part rimmed with veinlets of magnetite, and contain older laminae of a dark unidentifiable powder; crystals, needles, and grains of pleochroic yellow to brown piedmontite; and angular, fractured (?) quartz. The pyroxene (hedenbergite ?) is pale green, nonpleochroic, occasionally twinned; it surrounds the garnets. It is columnar with squat cross sections and contains lines of dust-gray powder. Muscovite occupies shears in the garnets. Sphene and calcite are ubiquitous. Quartz crosscuts all other minerals. Mineral percentages are: pyroxene, 50; garnet 40; the remainder, 10.

Sheafing through the garnet-pyroxene rock are veins composed of columnar zoisite (2V approximately 30°) with slightly oblique extinction, minor epidote, and clinozoisite.

For 30 feet above the contact, the limestone contains scattered garnet, sphene, chlorite, and muscovite. Clinozoisite (2V = 80°) wraps the garnets. Minor irregular quartz is also present; some detrital grains are suggested by their outlines. No wollastonite was detected.

Amphibolites

Immediately adjacent to the mountains, amphibolite separates diorite from limestone and schist (Hill, 1958a). The northern contact of the amphibolite, although poorly exposed, shows intermingling of amphibolite and diorite. The southern contact is faulted and dips steeply northward, with amphibolite resting above carbonaceous graphitic schist.

No extensions of the diorite and amphibolite were found within the mountains. To investigate the possibility that the amphibolite and the serpentinites might have a common magmatic source, seven samples were analyzed spectrographically for chromium, cobalt, and nickel. Semiquantitative comparisons were made with standard calibration powders (Table 5). More sampling and greater sensitivity are required before any significance can be attached to the similarity between results.

The origin of the amphibolite therefore remains uncertain. It may represent (1) an intrusion earlier than the diorite, (2) a basic border phase of the diorite (Thiadens, 1937), (3) limestone assimilation by a dioritic magma, (4) an intrusion later than the diorite, (5) metamorphosed diorite, or (6) metasomatized dolomitic limestone.

Deep weathering and soil cover prevent any definite solution to the problem. The meager outcrops that exist reveal gneissic amphibolite and diorite so intimately mingled that it is impossible to decide which is the intrusive body. A tentative explanation might involve theories 2 and 3. Later faulting has caused dynamic metamorphism of the amphibolite near the contact, but this does not explain the origin of the amphibolite, which averages 1200 feet in thickness.

The gneissosity and the apparent intrusive nature of the amphibolite, if not wholly developed when the diorite came into place, may have formed subsequently when the mountain front was flexed.

UNALTERED AMPHIBOLITE: The amphibolite where unaffected by cataclastic or dynamic metamorphism is a gneissic green and white rock. Its grain size approximates 1 by 5 mm. All samples are noticeably magnetic.

A hypidiomorphic-granular texture formed by hornblende, andesine (An_{45-55}), and accessory iron ores shows up under the microscope. The hornblende (2V = 65°) is strongly pleochroic from yellowish green (α) to blue green (γ). It is usually anhedral, but cross sections are subhedral.

The andesine forms large tabular crystals and smaller rounded ones. Both show lamellar albite and pericline twining.

The iron minerals are magnetite, ilmenite, and rare flakes of hematite. The ilmenite is partially altered to leucoxene.

ALTERED AMPHIBOLITE: Although the amphibolite appears to rest undisturbed on the underlying schist, locally it has been disturbed by fault movement. Dynamic metamorphism is exhibited in specimens 3–9 collected along the Amanayagua-Minas Carlota road (Pl. 4). Local metamorphism appears related to movement along several minor faults and shears. Therefore the degree of metamorphism, as seen in hand specimens, varies from sample to sample; altered but massive rock alternates with sheared schist. As expected, more advanced dynamic metamorphism is shown by the schistose varieties.

Thin sections of the massive rock reveal many changes. The hornblende has lost its pleochroism, and anthophyllite has developed sporadically along microshears (Pl. 3, fig. 8). The amount and crystal size of the andesine is decreased.

With advancing metamorphism new minerals appear. These include epidote (2V = 70° plus) wrapped by actinolite; pale-pink, broken garnets with separated fragments; quartz showing mortar structure or, where squeezed between amphibole and epidote, rare strain shadows; irregular stringers of sphene, surrounding ilmenite, which in turn surrounds deep-yellow rutile; and scattered, slightly biaxial apatite. Many sections of epidote, dark under crossed nicols, do not show cleavage and in ordinary light might be mistaken for olivine or diopside.

More advanced alteration in the massive rock has led to the disappearance of hornblende and andesine. Thin sections reveal porphyroblasts of clinozoisite set in a disturbed and cloudy groundmass (Pl. 3, fig. 9).

The clinozoisite (optically positive, 2V = 70°) is sheared and embayed by a groundmass of nephrite composed of fine antigorite and actinolite, nebulous and cloudy oligoclase-andesine (An_{20-35}), and sericite and chlorite replacing plagioclase. Sphene and larger laths of actinolite are scattered throughout.

The most disturbed rocks are the schistose

TABLE 5.—SEMIQUANTITATIVE SPECTROGRAPHIC ANALYSES OF IGNEOUS ROCKS
(Analyses by Mines Branch, Ottawa, Canada)

No.	Locality	Co	Cr	Ni	Nb (?)	B	Si	Mn	Mg	Fe	Al	V	Ca	Cu	Ag	Ti
1	Guaos Hill nodular serpentinite	0.01	0.4	0.5	0.03	0.004	10	0.2	15	5	1	0.01	0.03	0.007	0.01	0.03
2	Fine-grained serpentinite Wilson Ridge	0.025	0.13	0.38	..	..	..	..	..	..	..	..	..	..	..	..
3	Fine-grained serpentinite south of Gorro	0.007	0.03	0.04	..	..	..	..	..	..	..	..	..	..	..	..
4	Lawsonite-chlorite schist, gossan, 10 miles east of Charco Azul	0.015	0.02	0.05	..	..	..	..	..	..	..	..	..	..	..	..
5	Amphibolite, 1 mile northwest of Terminal Hill	0.006	0.01	0.01	..	..	..	..	..	..	..	..	..	..	..	..
6	Amphibolite, Locality 1 (Pl. 4)	0.006	0.07	0.06	..	..	..	..	..	..	..	..	..	..	..	..
7	Sheared amphibolite, Locality 9 (Pl. 4)	0.006	0.08	0.05	..	..	..	..	..	..	..	..	..	..	..	..

varieties. In thin section, laminae of fibrous actinolite are cut and coated by chlorite, talc, and possible nephrite; albite is riddled with needles of actinolite; individuals or masses of

grained variety (*see* gossans of Pl. 1) m perhaps be relicts of contact metamorphis The fine-grained type has been folded a faulted, whereas the nodular serpentinite a

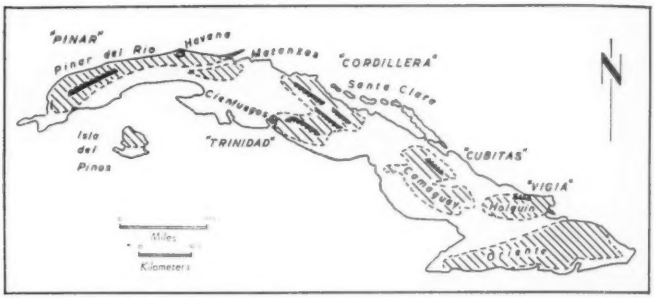

FIGURE 2.—POSTULATED TERTIARY THRUSTS IN THE STRUCTURAL HIGHS OF CUBA
(After M. T. Kozary, 1954, unpub. PhD thesis, Columbia Univ.)

zoisite, epidote, decomposed ilmenite, calcite, and sphene are ubiquitous. Quartz showing mosaic structure is a late introduction.

Problems of the Serpentinites

The division of serpentinites into fine-grained and nodular types is convenient for field distinction. The question arises, whether the two types are of the same origin, with different textures merely representing different tectonic environments.

Spectrographically, although the Guaos Hill nodular serpentinite and the Wilson Ridge fine-grained type show higher percentages of Co, Cr, and Ni than the other rocks analyzed (Table 5), the results are not conclusive in view of the small number of samples.

Microscopically the fine-grained serpentinites, compared with the nodular types, show gradation into, or association with, an actinolitic "greenschist" rock; occasional relicts of pyroxene; and, except for Honeymoon Hill, an absence of olivine pseudomorphs.

In the field, differences show up when the two types are compared. The nodular type, judging by its crosscutting outcrops (Pl. 1), is intrusive, and large masses of the fine-grained variety may also be intrusive. Smaller bodies of the fine-grained type apparently grade into epidote schists, suggesting that they may represent altered flows or tuffs. Although both types lack evidence of contact metamorphism, the pyrite deposits which surround the fine-

pears to have come into place after foldi (*e.g.*, Guaos Hill) and may thus be intrusi and younger.

Therefore mostly on field evidence, t fine-grained serpentinites are considered old than the nodular ones.

The origin of serpentinites is beyond t scope of this paper. Many hypotheses ha been advanced to account for their genesis a mode of emplacement (Hugh Gabrielse, 195 unpub. Ph.D thesis, Columbia Univ.). T relative ages of West Indian serpentinites a summarized by Mitchell (1955).

Although late faults are parallel to the co tacts between serpentinites and limestone no evidence exists that the serpentinites repr sent cold intrusions (Taliaferro, 1943; Kozar 1956). In addition nothing indicates th magnesian serpentinizing solutions were leache from dolomitic horizons. However shearin along possible tuffaceous horizons (*B* and faults, Fig. 3) may have formed chlorite schis by the addition of Al_2O_3 in the form of hydro aluminum silicates or hydrous alkali alumino silicates.

The concordant "bed ed" nature of many the serpentinites here described raises t possibility of a sedimentary origin. This p sibility may be eliminated, at least for t majority of the serpentinites, because th sections show that igneous rocks have be serpentinized, and the concentration of nick chromium, and cobalt is too great (Fau Murata, and Fahey, 1956).

FIGURE 3.—FAULT SYSTEMS OF THE NORTHWESTERN TRINIDAD MOUNTAINS
Age decreases alphabetically, *i.e.*, *A* is the oldest system

Structural Geology

Folding

Many of the isoclinal folds are bent. The bending may have been contemporaneous with isoclinal folding. However as three orogenies after the formation of the Trinidad Mountains are reported (Thiadens, 1937), the bending or cross folding may well have been later.

Faulting

Relief from isoclinal-folding pressure appears to have been predominantly "upward"; reverse faulting, much of it contemporaneous with isoclinal folding, is prominent. Most of these faults follow bedding and are difficult to recognize.

Tranverse faults, because of the lack of traceable units, are also difficult to recognize. Although the surface displacement of some of these faults is conjecturable, relative movement, whether normal, reverse, or tear (strike-slip), is unknown. So far as determinable, the intensity of folding on either side of these transverse faults is the same.

Interpretation of any faulted area consists of grouping the faults according to strike, noting whether one system offsets or is offset by another, and critically examining associated structures.

Difficulties in applying these criteria were at once apparent. Many faults change strike or may dip in opposite directions at different places. Many fault intersections are covered. Also, because one fault ends at, or is offset by another, it is not necessarily older—the two may be of the same age. Apart from the hard limonites of Gossans 2 and 3a (Fig. 3), the banding of which indicates that the underly ore bodies occupy tension cavities, no structu are preserved in the soluble limestone. schistosity, slickensiding, shearing, or joint were found along any fault.

Table 6.—Azimuths* and Tentative Ages of Faults

S.W. of Barona	Minas Carlota	North of Chivo	Amphibole Hill	South of Honeymoon Hill	Peak 5
E-W (B)	E-W (B)	NE-SW (A)	E-W (G)	E-W (G)	E-W (G)
NE-SW (C)	NE-SW (C)	E-W (B)	NW-SE (H)	NW-SE (H)	NW-SE (H
N-S (E)	N-S (E)	NE-SW (C)	E-W (I)	E-W (I)	E-W (I)
NW-SE (F)	E-W (G)	E-W (D)		N-S (J)	
E-W (G)	NW-SE (H)	N-S (E)			
NW-SE (H)					

* Azimuths are approximate; units vary up to 10° either side. Letters in parentheses correlate s tems; "A" is the oldest system.

Despite the limitations and difficulties volved, fault intersections and strikes w analyzed in certain localities to ascertain gene relationships (Fig. 3; Table 6).

Some generalizations appear tenable: movement along the east-west faults was peated; (2) system *E* may be used tentativ to divide the faults into younger and ol faults; (3) the oldest systems are in the inter of the mountains; (4) the northeast-southw systems (*A* and *C*) are among the olde (5) along the northern boundary of the mou tains, the majority of the *H* faults appear be late extension faults of "Appalachia type, diagonal to folding.

Structure of the Trinidad Mountains

The outer boundary of the Trinidad Mou tains—the contact between amphibolite a schist (or limestone)—appears to lack conta metamorphism. Thiadens (1937, p. 22) rep sented the contact as steeply faulted because the "... vertical and overturned position of t schists" and the "... strong cataclastic structu of the dioritic rocks".

Subsequently anomalies of Cuban structu have been explained as resulting from lar scale thrusts from north or south (Palm 1945; Thayer and Guild, 1947; Flint, Albe and Guild, 1948; Kozary, 1954, unpub. Ph thesis, Columbia Univ.; 1956; Wassal, 19 (Fig. 2).

Although the actual contact is not expos the demarcation is believed sharp. Gneissos

in the amphibolite parallels schistosity in the schist, and each rock was traced to within 15 feet of the other. Apart from large and sporadic actinolite crystals (along calcareous horizons ?) 300–400 feet from the contact, the schist is singularly undisturbed.

The present work therefore does not support northerly thrusting (Palmer, 1945). In the area studied the contact appears to be a high-angle reverse fault.

Later transverse faults displace the contact. Some, *e.g.*, west of Terminal Hill, form scarps between rocks normally of equal resistance to tropical weathering. These may be recent. Keijzer (1944) describes Pleistocene faulting from other parts of the island.

The extent to which thrusting exists within the mountains is debatable. Field conditions are not good. The proportion of cover to rock 8:1. Critical outcrops may be hidden by forest and overburden. However Charles Hatten (1958, personal communication) writes:

"To my knowledge, the best area within the main Sierra de Trinidad where thrusting is evidenced is in the eastern end of the mountains (Sierra de Sancti Spiritus). East of the Rio Bañao the general attitude of the metamorphics (quartz mica schists with minor interbedded calcareous schists) is north-south with a 60–80° east dip. Minor drag folds and poorly developed fracture cleavage indicate that the sequence is right-side up. To the west of Rio Bañao a sequence, obviously different lithologically from the above, with calcareous schists predominating, also strikes north-south but dips steeply to the west. Near the base of this latter sequence, cataclastites are well developed. Minor recumbent isoclinal folds are present, which I believe, plunge to the northwest. The structural discordance between these two units is interpreted as a westward dipping thrust fault; however it could be an unconformity. More work should be done in this area to determine the relationship."

To summarize; in the area studied, the mountains are not overthrust from the south. As suggested by Thiadens (1937) they appear to form a gigantic anticlinorium.

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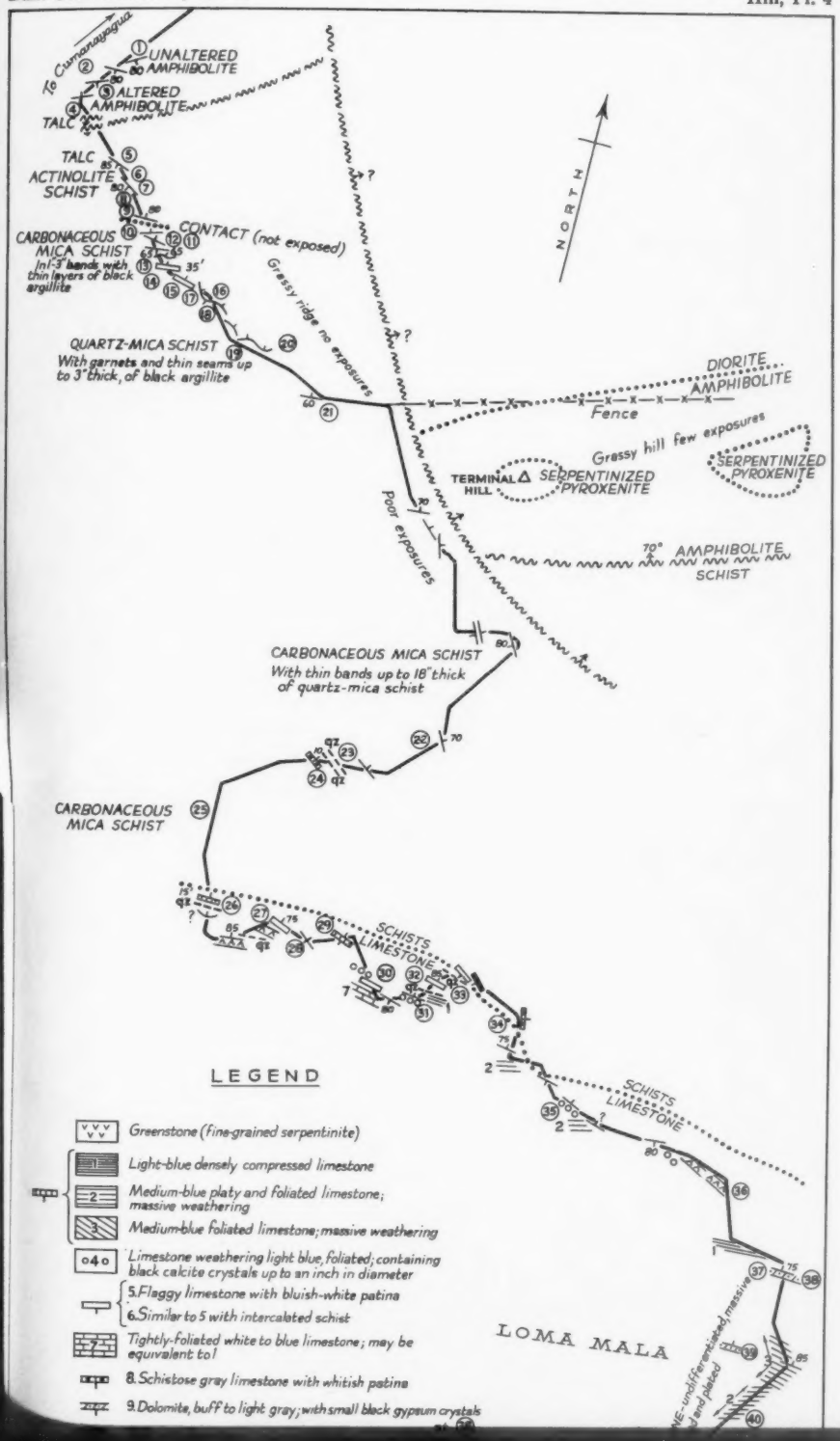
Cumanayagua
1 UNALTERED AMPHIBOLITE
2
3 ALTERED AMPHIBOLITE
4 TALC
5 TALC ACTINOLITE SCHIST
6
7
8
9
10 CONTACT (not exposed)
CARBONACEOUS MICA SCHIST
In 1-3' bands with thin layers of black argillite
11
12
13
14
15
16
17
18
Grassy ridge no exposures
QUARTZ-MICA SCHIST
With garnets and thin seams up to 3" thick, of black argillite
19
20
21
NORTH
DIORITE AMPHIBOLITE
Fence
Grassy hill few exposures
TERMINAL HILL
SERPENTINIZED PYROXENITE
SERPENTINIZED PYROXENITE
Poor exposures
70° AMPHIBOLITE SCHIST
CARBONACEOUS MICA SCHIST
With thin bands up to 18" thick of quartz-mica schist
22
23
24
CARBONACEOUS MICA SCHIST
25
26
27
28
29
SCHISTS
LIMESTONE
30
31
32
33
34
35
SCHISTS
LIMESTONE
36
37
38
39
40
LOMA MALA
LEGEND
Greenstone (fine-grained serpentinite)
1 Light-blue densely compressed limestone
2 Medium-blue platy and foliated limestone; massive weathering
Medium-blue foliated limestone; massive weathering
Limestone weathering light blue, foliated; containing black calcite crystals up to an inch in diameter
5 Flaggy limestone with bluish-white patina
6 Similar to 5 with intercalated schist
Tightly-foliated white to blue limestone; may be equivalent to 1
8 Schistose gray limestone with whitish patina
9 Dolomite, buff to light gray; with small black gypsum crystals

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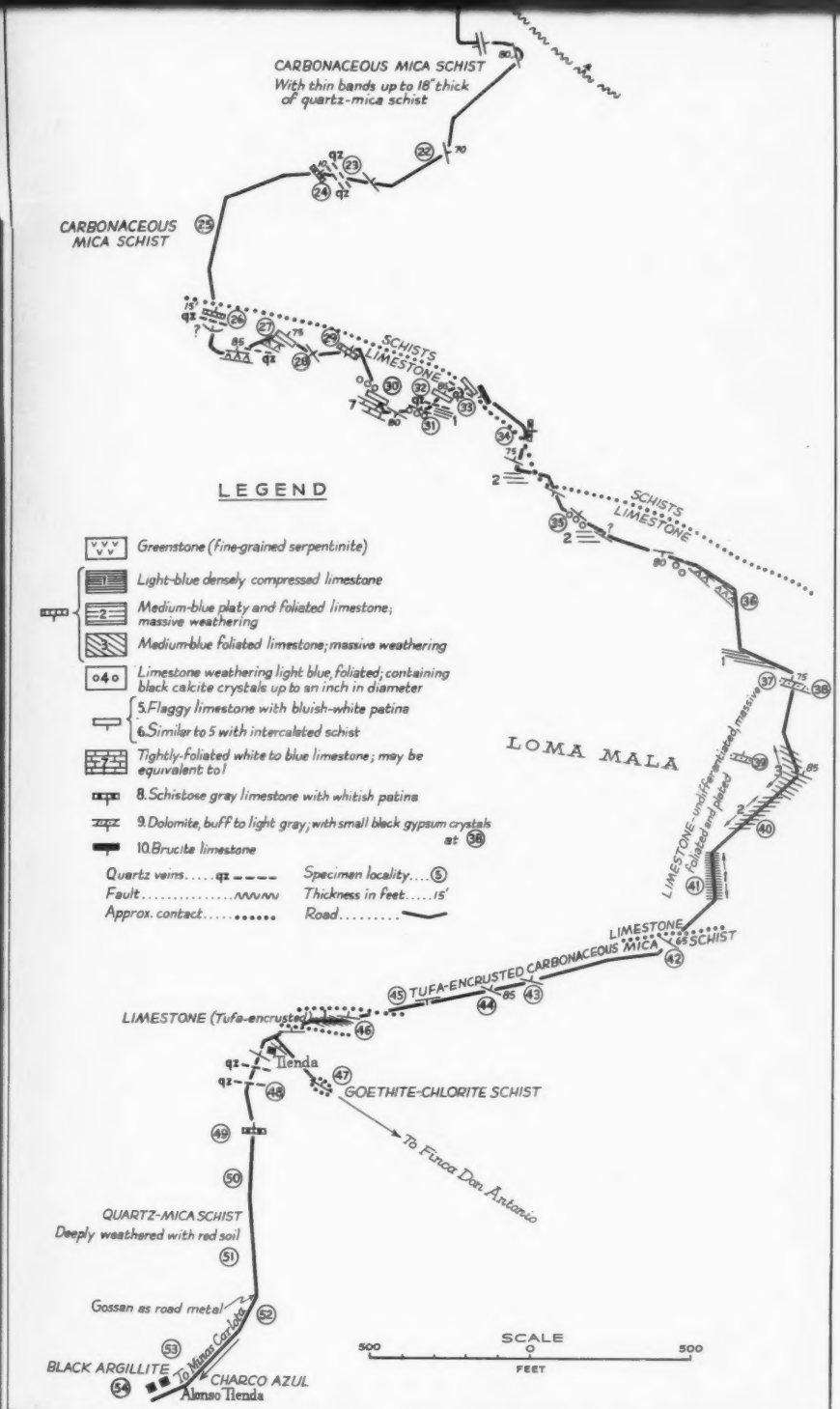

GEOLOGIC TRAVERSE ALONG THE CUMANAYAGUA-MINAS CARLOTA ROAD

From Terminal Hill to Charco Azul

Short Notes

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
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SLOPE RETREAT BY GULLYING

BY CHESTER B. BEATY

The mechanisms of slope retreat are varied, and few generalizations concerning this fundamental gradational process are valid for more than limited areas. Climate, slope, vegetational cover, thickness of regolith, and lithologic character of underlying bedrock are the major factors which control the rate at which a slope retreats and declines in steepness as well as the processes of recession active thereon. Radical changes in the controlling factors even in short horizontal distances make it impossible to generalize concerning slope processes.

An appreciation of processes of slope retreat is basic to an understanding of land-form evolution and alteration. Yet all too often the importance of slope processes is ignored or only perfunctorily treated. Long ago William Morris Davis called attention to the significance of denudational activities on canyon walls, as opposed to the work of rivers on their floors. In *The Geographical cycle*, Davis (1899, p. 266) stated that

"... valleys are eroded by their rivers, yet there is a vast amount of work performed in the erosion of valleys in which rivers have no part. It is true that rivers deepen the valleys in the youth and widen the valley floors during the maturity and old age of a cycle, and that they carry to the sea the waste denuded from the land; it is this work of transportation to the sea that is peculiarly the function of rivers, but the material to be transported is supplied chiefly by the action of the weather on the steeper consequent slopes and on the valley sides. The transportation of the weathered material from its source to the stream in the valley bottom is the work of various slow-acting processes, such as the surface wash of rain, the action of ground water, changes of temperature, freezing and thawing, chemical disintegration and hydration, the growth of plant roots, the activities of burrowing animals. All these cause the weathered rock waste to wash and creep slowly downhill..."

Somewhat later—1922—the importance of mass movement in slope reduction was stressed by Walther Penck; mass movement was defined as

"... the spontaneous independent migration of the reduced substances ... on all slopes where rocks are being reduced ... where the inclination exceeds a minimum amount which observation shows to be removed by only a few degrees from the horizontal" (Penck, 1953, p. 74).

Still later a detailed discussion and classification of various manifestations of mass movement was provided by C. F. S. Sharpe's *Landslides and related phenomena* (1938).

More recently, specific slope studies have revealed a variety of processes by which debris is moved downslope toward and into organized channel systems (Horton, 1945; Strahler, 1952; Schumm, 1954; Strahler, 1956). One such process is slope gullying. Under the right environmental conditions, gullying can be a most effective means by which slopes are lowered and altered.

This study reports the results of an examination of gullying in the White Mountains of California and Nevada (Fig. 1). The investigation was made in connection with a study of desert flooding for the U. S. Army Quartermaster Corps (Kesseli and Beaty, 1959).

The White Mountains are the highest range in the southwestern Great Basin, lying directly east of the Sierra Nevada across Owens Valley, California. Elevations of the crest range from 10,000 to 13,000 feet, and White Mountain Peak, culminating summit, attains a height of 14,246 feet, only 250 feet lower than Mt. Whitney. Adjacent valley floors are 4000 to 6000 feet in elevation; thus local relief in the range is great, stream gradients are high, and slopes are notably steep, especially along the western front. Climatically the range is Cold Steppe—BSk—along its crest, whereas flanking valleys are Cold Desert, BWk. Precipitation is normally concentrated in the winter, falling from frontal storm systems moving from the Pacific Ocean. But the sporadic summer thunderstorm is a major agent of morphologic change also, and produces flash floods and

debris flows in the canyons and on contiguous alluvial fans, as well as gullying on favorable slope segments. In so far as sudden and drastic physiographic effects are concerned, the summertime cloudburst is the single most significant element of the climate of the area.

Many White Mountain slopes are furrowed by systems of parallel or semiparallel gullies which are obviously water cut. Such slopes appear to be retreating at the time of the infrequent heavy rain by a process analogous to what was termed "gully gravure" by Kirk Bryan (1940). The essential feature of Bryan's gully gravure is the lateral transposition in time of gullies and ridges. In his words:

> "Gullies in the slope are gradually filled with detritus which forms a pervious mantle, absorbing run-off and inducing further deposition. In the meantime the inter-gully ridges decay and become the locus of run-off so that the softened rock and detritus from a previous cycle are removed to form a new set of gullies. The previous gullies capped by detritus are now ridges, subject to decay and reversal in position" (Bryan, 1940, p. 105).

It is difficult to accept Bryan's contention that the perviousness and imperviousness

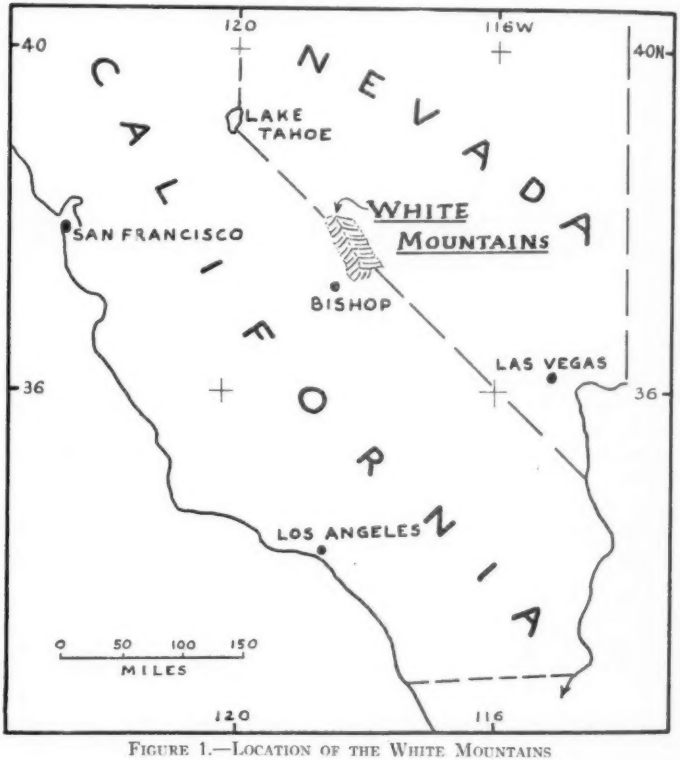

FIGURE 1.—LOCATION OF THE WHITE MOUNTAINS

adjacent strips of hillside would be so marked that a reversal of position of gullies and ridges could occur. Running water seeks and follows depressions; the effect is to deepen existing channels, not to create new ones where once stood ridges of higher ground.

On lower White Mountain slopes smooth segments are being carved by gullies, and the slopes are retreating essentially by a process of "gullying." But there is little evidence to suggest that ridges and gullies change position cyclically. The inter-gully ridges are bedrock thinly veneered with weathered debris, not "gullies capped with detritus." Tiny rill systems and minute debris lobes on the sides of the ridges indicate that the loosened material moves off the ridges and into adjacent gullies. The floors of the gullies exhibit all degrees of

debris accumulation, from deposits 1–2 feet thick on those which have not been flushed out by surface runoff for a long time, to only traces of sand and silt on the bedrock bottoms of those which have recently been cleared. Evidence suggests that existing gullies continue to deepen, and intergully ridges are lowered at a roughly comparable rate (Fig. 2).

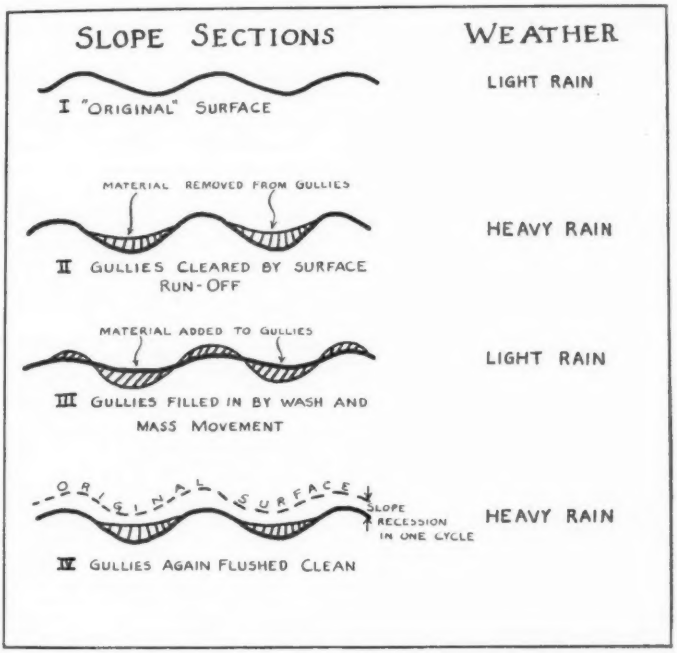

FIGURE 2.—DIAGRAMMATIC REPRESENTATION OF SLOPE RETREAT BY GULLYING
No set time interval between heavy rains is specified.

During heavy rain, water collects in the gullies and flushes out weathered debris accumulated from adjacent ridges. Gully floors are also deepened by scour, and undoubtedly a certain amount of ridge lowering goes on as a result of minor rill cutting and washing. Also, material moves from ridge tops and sides toward gully floors in the form of small debris lobes or tongues, small-scale replicas of larger landslides in more humid regions. But the dominant process during periods of torrential precipitation is gully deepening.

During the relatively long periods between really heavy rains, weathering loosens material on the ridges and in the gullies and prepares it for transportation. During and following light rains, and especially during periods of soil saturation brought about by snow melt, mass movement, in the form of the debris lobes, and rill cutting shift material into gully floors. These processes lower the ridges. Gullies are places of accumulation, ridges places of removal. The slope is smoothed, and the total morphologic effect of a completed cycle is a lowering of the whole slope surface. When heavy rain falls, surface runoff again collects in the partially filled gullies and again flushes them out, thus intensifying local relief. The cycle is repeated; slice by slice the slope is literally peeled away. The process appears to be most active on metamorphic and sedimentary rocks covered with a thrifty growth of shrubs and grasses, but gullies were also seen on slopes underlain by granite and basalt.

Some of the material moved downslope by the gullying goes directly into integrated channel systems in which further transport is provided by perennial streams. However, most of the debris swept out of the gullies comes to rest on lower, more gentle slopes, where processes of mass movement assume importance in its subsequent downhill trans-

fer. Detritus delivered to the upper ends of the gullies by mass movement and rill cutting on segments above the furrowed facets becomes part of the layer of loosened material to be moved by the next heavy rain. The gullying process, in short, is not an isolated phenomenon.

Lateral shifting of the gullies could well be brought about by differences in bedrock resistance, but probably the initial surface pattern of furrows is superimposed indiscriminantly upon the underlying structure, and a parallel or subparallel drainage system prevails throughout the life of the slope. All the gully systems observed had a parallel or subparallel arrangement of the furrows, and they were seen on sleep slopes (35°) as well as on hillsides with an inclination of 10° or less. In any event, this is slope retreat by gullying *par excellence* and is one of the more important processes by which some range-front and canyon-wall slopes of the White Mountains are being consumed.

Ridge and gully systems were found on slopes with an average inclination of 15°–20°. The gullies examined in detail ranged in length from 25 to 200 feet, in width from 1 to 3–4 feet, and in depth from 4 inches to 3 feet. There was no observable correlation between size of gully and type of underlying bedrock, but the longer and deeper gullies are more common on the steeper slopes.

Gullying on White Mountain slopes is but one of a series of interconnected processes by which loosened surface debris is moved downslope. Gravity fall, mass movement, and rill cutting further the efforts of gullying, and all constitute the denudational processes by which slopes in the range are altered. Gullying appears to be favored by a combination of comparatively smooth slopes with an inclination of 15°–25° on which the vegetation cover is sparse and composed of grass or shrubs. In an area subject to infrequent but heavy rains, such slopes may retreat dominantly by the gullying process. Since the process was observed on slopes underlain by metamorphic, sedimentary, and igneous rocks, type of bedrock cannot be of primary importance.

Gullying in the White Mountains is a sporadic process, depending mainly upon the occurrence of torrential rain. Recession of the slopes thus goes on at an irregular rate. Any slope segment may remain inactive for years or even decades. But eventually the cloudburst hits, and decisive slope alteration is accomplished. In the irregular periodicity of its occurrence, slope gullying resembles other processes of gradation in the desert environment.

But it should be stressed again that gullying is only *one* process at work; the results of this process are undeniably visually spectacular, but perhaps no more important in the long run than are those of the other mechanisms of slope retreat.

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GUMBOTIL AND INTERGLACIAL CLAYS

BY VICTOR T. ALLEN

Lougee's *Ice age history* (1958a; 1958b) introduces several geological problems which geologists should examine critically. These include the deposition of erratics and tills from "iceberg-rafted erratic stones and boulders" (1958a, p. 1291).

My discussion is directed mainly toward Lougee's concept of gumbo clays and the relationship of these clays to his thesis that "deposits formerly attributed to four or five separate Pleistocene glaciations, both in America and in Europe, are deposits of a single glaciation" (1958a, p. 1290). Lougee's concept of gumbo clays is contained in two quotations:

"Gumbo clays, until recently interpreted to be weathered tills, were deposited within the expanse of the sea-level waters, along with drift-wood and other organic material heretofore interpreted to be 'interglacial deposits'" (1958a, p. 1291); "Directly overlying the old tills there are widespread clays and silts, variously known as 'Loveland or Farmdale', or 'gumbo,' which may be weathered phases of a deep-water deposit" (1958b, p. 1764).

Petrographic Study of Weathered Zones of Glacial Deposits in Illinois and Iowa

Under the auspices of the Illinois State Geological Survey 10 profiles of weathering developed on glacial deposits of Wisconsin, Illinoian, and Kansan glacial stages were studied petrographically (Allen, 1930) in connection with the field studies by Leighton and MacClintock (1930) on the weathered zones of the drift sheets of Illinois. Complete oriented vertical sections of the weathered zones were cut in the field at selected localities and placed in tin trays for laboratory investigation. The thin sections prepared at that time are still available and were restudied during preparation of this discussion.

Leighton and MacClintock (1930, p. 31) recognized the following subdivisions of the weathered zones developed on the glacial drift sheets in Illinois:

Horizon 1 : the surficial soil
Horizon 2 : chemically decomposed till, composed chiefly of alteration products and resistant constituents of the original till
Horizon 3 : leached and oxidized till, otherwise but little altered
Horizon 4 : oxidized till, but unleached and otherwise unaltered
Horizon 5 : unaltered till

Petrographic study of the weathered zones developed on the Wisconsin, Illinoian, and Kansan drift sheets shows that Horizon 5, unaltered till, consists of a heterogeneous mixture of clay and rock fragments varying in composition with the source rocks. Horizon 4 is oxidized and unleached till, but otherwise it is similar to Horizon 5. The characteristic features of Horizon 4 are its staining with yellow and brown iron oxides and the presence of fragments of limestone and dolomite. Samples of Wisconsin and Illinoian till that were immersed in 0.1 normal HCl lost about 20 per cent by weight, and this furnishes an approximation of the percentage of carbonates in the samples examined. Horizon 3 lacks carbonates, and in the upper part there is a transitional change to Horizon 2. In Horizons 2 and 1 evidence of chemical decomposition *in situ* is afforded by the edges of feldspar and ferromagnesian grains which are so intricately altered that it is difficult to decide where the edges of the feldspar leave off and where the clay begins. Under crossed nicols veinlets and patches of clay divide some grains into several parts that together extinguish as a single grain. Quantitative measurements with a Wentworth recording micrometer on the particles in thin sections more than .02 mm across show that there is less feldspar in Horizon 1 and 2 than in the lower horizons. Furthermore, the remaining feldspar grains are smaller and consist of the more resistant members, microcline, perthite, and albite. The processes that operated were selective; they removed some minerals such as orthoclase and plagioclase that are abundant in the unaltered till, and partially altered and reduced the grain size of others. Ferromagnesian minerals are similarly affected. Numerous grains of hornblende and epidote have been observed altered to clay minerals at the edges and along cleavage cracks.

In all the profiles investigated the percentage of heavy minerals decreases toward the top. Comparisons of weighed amounts of heavy minerals separated from equal-weight samples of Horizon 2 and Horizon 4 showed a one- to

fivefold decrease in the amount of heavy minerals toward the top of the profile. Reductions in amount are limited to the minerals which perish easily, such as apatite, magnetite, hornblende, and epidote. The minerals formed during the destructive process include clay minerals of the montmorillonite group that contain notable amounts of the iron member nontronite, as well as iron oxides.

Leighton and MacClintock (1930, p. 39–46) recognized variations in the types of profiles developed that are related to topographic situations and conditions of drainage. In poorly drained areas Horizon 2 is composed of a plastic clayey material named gumbotil by Kay (1916). Petrographic studies indicate that gumbotil is composed chiefly of clay minerals of the montmorillonite group in which the iron member nontronite is notably represented. In this clay are embedded grains of quartz, sandstone, chert, and other resistant minerals and siliceous rocks and a few grains of partly altered feldspars that increase in amount and decrease in the degree of alteration progressively downward. Many feldspar grains in gumbotil extinguish as a single grain but are so cut by veins and patches of clay minerals that they could not have been transported in this condition. Petrographic evidence strongly supports Kay's conclusion that gumbotil is essentially the result of chemical weathering of glacial till (Kay and Pearce, 1920, p. 124). In well-drained areas, a friable silt forms Horizon 2; in partially drained areas Horizon 2 is less friable. For the former Leighton and MacClintock have proposed the term silttil; for the latter, mesotil. In mesotil and silttil evidence is convincing of the alteration *in situ* of feldspars to clay minerals. However, the clay minerals formed in Horizon 2 do not remain *in situ* as they do under conditions of poor drainage to give plasticity to this horizon, but migrate downward. In Horizon 3 are cracks and cavities filled with clay minerals that have migrated from above. The cleavage plates of the clay minerals are oriented parallel to the sides of the cracks, and successive layers vary in indices of refraction and in content of nontronite. This suggests secondary deposition. Downward movement has probably taken place as colloidal solutions because of the large quantity of water passing through the profile under conditions of good drainage. By repeated washing clay becomes so finely divided it will pass through a quantitative filter paper. Particles of clay that were washed through a filter paper in the laboratory oriented themselves on a glass slide into a continuous sheet that resembles the cleavage sheets filling cracks in the field exposures.

The second-cycle profiles of Leighton and MacClintock are formed where erosion works headward into a region with poor drainage and brings about improved drainage conditions. Horizon 2 is changed to mesotil or silttil, and Horizon 3 shows the orientated plates of montmorillonite that are present in the profiles of partially and well-drained areas. This is further evidence of the modifying effect of subsurface drainage on the distribution of clay minerals in the profiles.

Petrographic Studies of Residual Clays Formed by Chemical Weathering in Situ

The bedrock below the Ione formation of California and the quartz gravels of Eocene age that lie on the slopes of the Sierra Nevada is altered to residual clays containing angular quartz grains, euhedral zircon, anauxite, and other clay minerals. Petrographic study of outcrops and cores (Allen, 1929, p. 387–388; 1946, p. 125–126) indicates that orthoclase, plagioclase, biotite, hornblende, sphene, and magnetite, which are present in the granodiorite of the Sierra Nevada, are removed by chemical weathering from the upper part of the zone of weathering, leaving chemically resistant angular quartz grains and euhedral zircons in a matrix of clay minerals. At depth feldspars and ferromagnesian minerals are present, and their edges are altered and cleavages invaded by clay minerals. The alteration is greatest near the surface and decreases downward where partly altered minerals indicate the stages in their progressive decomposition *in situ*.

Thin sections of cores and outcrops in the Columbia River region (Allen, 1948) record the selective decomposition of less resistant minerals throughout a thickness of 175+ feet of basaltic rocks. The calcic core of plagioclase is altered to nontronite, whereas the sodic rim remains unaltered. Augite weathers to nontronite, and plagioclase alongside remains fresh (Allen and Scheid, 1946, p. 305–310). Veins and cracks are filled with nontronite that migrated through the clay under conditions of good drainage.

Petrographic Studies of Sedimentary Clays

The writer has studied Pleistocene clays in New England and Eastern Canada. The clay minerals of varved clays and marine clays that were studied from this region are illite or

hydrous mica and chlorite or vermiculite (Allen and Johns, manuscript). These are the same at the top and at the bottom of exposures from which samples were collected. Petrographic study of thin sections of these fresh sedimentary clays indicated the lack of feldspars or ferromagnesian mineral fragments with edges more altered than the interior. Evidence is also lacking for the presence of chemically resistant minerals at the top of deposits and the progressive increase downward in the amount of perishable minerals.

Discussion

The processes suggested by Lougee (1958a, 1958b; 1958c) for the origin of till will not explain the features of the weathered zones of glacial deposits or of gumbotil. To account for large fragments in a water-laid clay he advocates having them dropped from icebergs into the water. Not only would the iceberg have to select the required size of particle at every instant but also to sort out the minerals so as to deposit the most resistant minerals near the top and increase the amount of perishable minerals progressively downward. Even if it were possible to do this, the deposition of grains of feldspars and ferromagnesian minerals with their edges intricately altered to clay minerals could not be done without wearing the clay from the edges. Nor could grains cut by numerous clay veins be transported in water without coming apart. The petrographic relations indicate decomposition of minerals *in situ*.

The writer also disagrees with Lougee's statement:

"It appears that Kansan and Illinoian tills are the same in age and degree of weathering although derived from different ice centers; that they are deposits of one glaciation only" (1958b).

My petrographic study indicates that the weathered zones on the Kansan drift at Fairfield, Iowa, are thicker and show more intense weathering than do those formed on the Illinoian drift near Salem, Helm, and Effingham, Illinois. These observations support the conclusion of Kay (1931) that depth of leaching can be used to estimate the duration of the interglacial stages.

Conclusions

The origin of weathered zones on glacial drift sheets of Illinois and Iowa by the dropping of "ice-rafted stones" into water-deposited clay is not in harmony with observations at classical localities in Iowa and Illinois. Petrographic study of gumbotil shows the edges of feldspar grains altered to clay minerals and shows grains cut by clay veins that could not be transported in water without wearing the clay from the edges and the separation of the grains into parts. Nonresistant minerals have been removed from the upper part of the weathered zone by chemical weathering *in situ* but increase in amount progressively downward. Where weathered drift is overlain by unweathered drift a long period of weathering separates the glacial deposits. The presence of gumbotil is proof of a long interglacial interval. Silttil and mesotil grading downward into a zone containing veins and cracks filled with clay minerals also indicate a long period of exposure of the drift sheet on which each is developed.

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VERTICAL PHOTOGRAPH OF CENTRAL PART OF MODEL PIEDMONT GLACIER

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SERPENTINE MEDIAL MORAINES ON MODEL GLACIER

BY W. T. MCCUTCHEN AND W. F. TANNER

Sharp (1958) described folds found in Malaspina glacier, on the southern border between Alaska and Canada. Those folds were not inherited from the Seward glacier system, which feeds the Malaspina, but have their origin in the movements of the Malaspina. After a study of the map patterns formed by the medial moraines involved, and the fracture network, Sharp concluded that the origin of the folds is still a puzzle.

The present note describes, briefly, a model designed to duplicate the serpentine pattern of the medial moraines on the Malaspina. The following model ratios were used (Hubbert, 1937):

Length	1.42×10^{-5}
Gravity	1
Density	About 1
Strength	About 1.42×10^{-5}

The density factor was actually close to 1.3, inasmuch as clay mud was used as the model material. A possible temperature ratio was excluded because it was thought that the folds have a mechanical origin. One hour of laboratory operation was considered equal to, roughly, 20 years of glacial movement.

The model was contained in a box 50 cm square. Only half the glacier was modelled, inasmuch as it was thought that the other half was not necessary for the purpose at hand. The "rock basin" was formed from water-soaked newspaper and covered with aluminum foil to prevent water loss from the clay. The clay was about 2 cm thick.

Three mud reservoirs were constructed to represent three of the tributary glaciers. The central reservoir was supplied with a white kaolin mud, whereas gray-green muds were used in the two outer reservoirs. The sharp color contrast permitted detailed study of relative motion in the "medial moraine" or boundary zones. The boundaries between the several muds were vertical or very steep.

The working hypothesis was adopted that differential flow from multiple reservoirs was responsible for the formation of the serpentine pattern. Differential flow was achieved by feeding the reservoirs at different rates. Six "runs" were made, three of them with differential flow; two of the latter, and none of the others, produced the serpentine pattern. Each "run" was continued for about 1 hour.

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MANUSCRIPT RECEIVED BY THE SECRETARY OF THE SOCIETY, MAY 22, 1959

PLATE 1.—VERTICAL PHOTOGRAPH OF CENTRAL PART OF MODEL PIEDMONT GLACIER

The three reservoirs, which supplied dark, light, and dark mud, respectively, are located above the top of the picture. The flow was toward the bottom of the picture. The fracture network has been inked at selected points. The finely digitate pattern between the two colors of mud (for example, just to the right of the dime) should be noted. The high shadow contrast was provided by a very low light source. (Dime included for scale)

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 70, PP. 1489–1490, 1 FIG., 1 PL. NOVEMBER 1959

FORMATION OF VENTIFACTS IN A MOIST, TEMPERATE CLIMATE

BY CHARLES F. HICKOX, JR.

Ventifacts have been found in the central portion of the Annapolis Valley, Nova Scotia. Field relationships indicate that these ventifacts have been faceted under the existing climate and within the last 10 years.

The average annual rainfall for the Annapolis Valley is about 41.41 inches. This figure is based on cumulative statistics from the weather station at Annapolis Royal for a period of 25 years (Climatic Summaries, p. 54).

No statistics of wind velocity are available for the Annapolis Valley. The prevailing wind, however, is westerly and of moderate velocity, similar in strength and persistence to the prevailing westerlies of coastal New England and southeastern Canada. The Annapolis Valley, therefore, has moist, temperate climate.

The ventifacts of the Annapolis Valley are restricted to the surface of an esker delta. This ridge of ice-contact stratified drift, 8.0 miles long, trends southwestward across the Annapolis Valley from Aylesford to Melvern Square. North of the village of Kingston the ridge is 0.9 mile wide and rises about 30 feet above onlapping sediments of an outwash plain.

The internal structure of the delta, exposed in a series of sand pits, consists of cross-laminated beds of well-sorted, medium-grained sand with rare lenses of water-worn pebbles that range from subrounded to subspherical. Most of the pebble-sized particles are pink quartzite, white vein quartz, or pink sandstone. The remaining pebbles are granite, slate, or basalt.

The highest and widest portion of the ridge, north of Kingston, is very well drained because of the permeable sediments and the relief of the ridge. During normal summer weather the water table, along the axis of the ridge, is about 25 feet below the surface. Because of the excessive drainage about one-third of the surface of the ridge is unvegetated sand; the remaining two-thirds is blanketed by blueberry bushes, heath plants, or open pine forest. The unvegetated patches along the crest of the ridge are being deflated, and the finer sand has been winnowed from the surface leaving a lag concentrate of coarser sand and scattered pebbles.

The pebbles have been modified from their original subrounded or subspherical water-worn form into faceted and keeled ventifacts; vein quartz, pink quartzite, and indurated sandstone have been abraded into ein-, zwei-, and dreikanters; less homogeneous granite and slate have been pitted or abraded about softer mineral grains or along zones of low resistance (Pl. 1, fig. 1).

The relationships—topographic relief, permeable sediments, deep water table, unvegetated surface, lag concentrate, and wind-abraded pebbles—are significant because surface conditions of the ice-contact delta are similar to those of an arid environment even though the local climate is moist. Because the ventifacts are surficial it is inferred that they have been abraded in modern times and are still being abraded, under existing climatic conditions.

A sand pit, excavated in the crest of the esker delta, 0.7 mile north of Kingston post office, furnishes a clue to the time required for the formation of ventifacts. Ventifacts, found in association with the sand pit, were first studied during the summer of 1955. The pit, however, does not appear on air photographs taken in the spring of 1945. Furthermore the owner of the pit, Mr. Gilbert Hudgins, states that the first sand hauled from the pit was used in construction of a concrete hangar apron at Greenwood Air Base. This apron was not completed until after the Nazi defeat in May 1945. The pit, therefore, must have been excavated between 1945 and 1955.

During excavation of Hudgins' pit, now 160 feet in diameter and 25 feet deep, mounds of sand up to 5 feet high were backfilled on both western (windward) and eastern (leeward) sides. These mounds have since been modified into dunes by the prevailing westerly wind. The fine and medium sand has been winnowed from the windward face and has been deposited on the leeward slope where it assumes an angle of repose of 27°. Coarse sand and pebble gravel remain as lag concentrate on the gentle windward slope of the backfill (Pl. 1, fig. 2).

The pebbles on the windward slopes of the backfill, like those on the unvegetated portions of the ridge, have been abraded by the wind

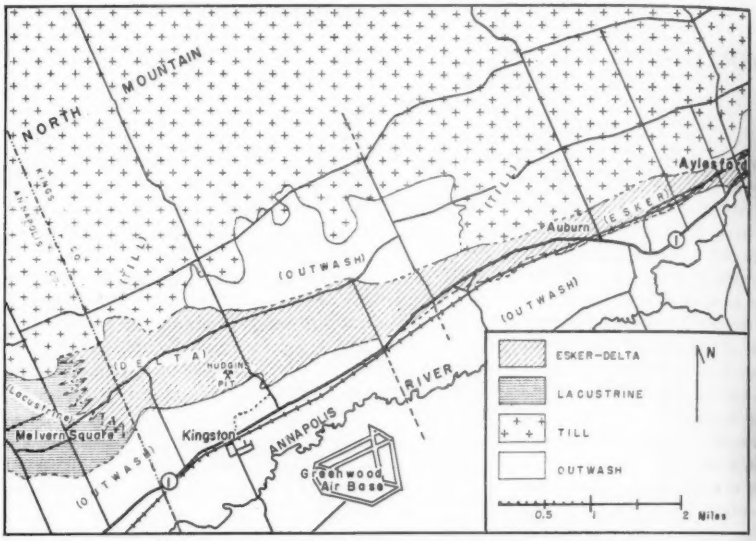

FIGURE 1.—MAP OF CENTRAL PORTION OF ANNAPOLIS VALLEY, NOVA SCOTIA

into ein-, zwei-, and dreikanters or have been pitted or abraded about softer minerals (Pl. 1, fig. 1).

Ventifacts occur on the windward slopes of backfill on both east and west sides of the sand pit. Therefore the unvegetated terrain upwind from the pit, rather than the pit itself, is the source of the abrading wind-blown sand.

Thus, ein-, zwei-, and dreikanters can be cut on vein quartz, quartzite, and indurated sandstone in less than 10 years. Prevailing winds of moderate strength and a windward source of dry, transportable sand are the only requirements for abrasion of ventifacts. Ventifacts can form in a moist, temperate climate; they do not necessarily indicate arid or proglacial environments.

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PLATE 1.—VENTIFACTS FROM ANNAPOLIS VALLEY, NOVA SCOTIA

FIGURE 1.—Water-worn pebbles collected from gravel lens, Hudgins' Pit (top row) and ventifacts collected from windward slope of backfill (bottom row)

FIGURE 2.—Backfill modified into dune, Hudgins' Pit

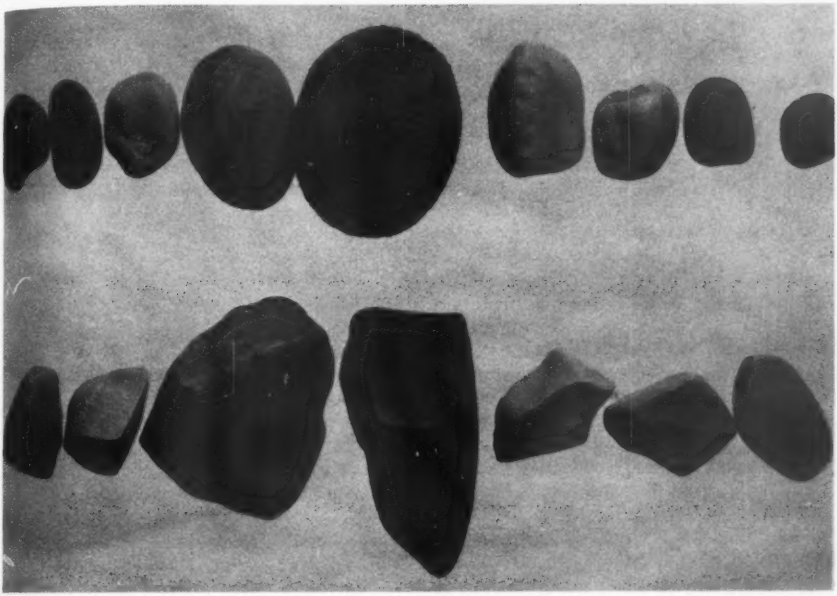

FIGURE 1

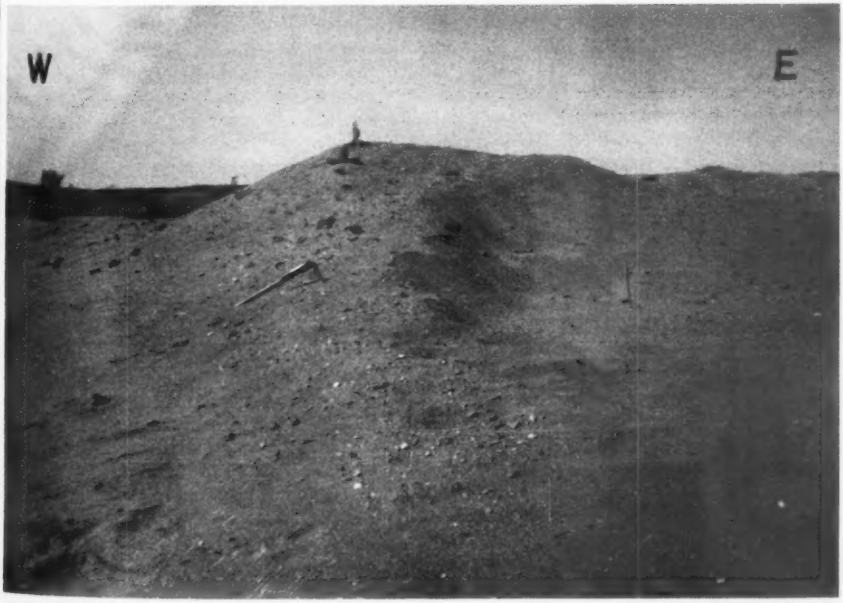

FIGURE 2

VENTIFACTS FROM ANNAPOLIS VALLEY, NOVA SCOTIA

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
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MINOR LINEATIONS IN THE PACIFIC BASIN

BY H. W. MENARD

Widely distributed minor lineations in the cific Basin lie in a pinnate pattern relative the major lineations such as island arcs, chipelagoes, and fracture zones. Individual inor lineations generally are 20–200 km long d consist of: (1) elongate submarine vol-noes, atolls on volcanic platforms, and vol-nic islands; (2) lines of closely spaced circular bmarine volcanoes; (3) asymmetrical ridges;) narrow troughs ordinarily associated with rallel asymmetrical ridges; (5) straight roughs resembling submarine canyons on con-ental and insular slopes.

Compared to faults on land these "minor" eations are rather large features, but at sea e most detailed techniques of location and ry closely spaced sounding lines are required survey most of them. Within a few hundred iles of land, electronic positioning is possible. rther out to sea, buoys are planted, or an tricate pattern of dead reckoning may be ed. The requirement of accurate positioning no more critical than accurate sounding with cording, high-performance echo sounders. r this reason the large network of random ot soundings in the Pacific is of little use in tecting minor lineations. Narrow troughs are rticularly elusive because the steep sides all t disappear on the record, and it may appear at little change in depth has occurred on the ased type of recorder used by oceanographers. lmost 200 minor lineations are known in the acific Basin at present. Inasmuch as all of em are in specially surveyed areas which mount to only a small fraction of the basin ea, it may be anticipated that thousands main to be found.

Elongate volcanoes are the most common inor lineations in the Pacific Basin. Excellent amples are San Juan Seamount (Shepard nd Emery, 1941, Pl. 4), San Benedicto Island the Revillagigedo Group (Richards, 1957, heses, Univ. Calif. at La Jolla, Pl. 2A), and akaroa Atoll in the Tuamotu Archipelago. ndividual elongate volcanoes grade into rcular volcanoes which are so closely spaced at they have a common base. Clipperton land and the adjacent seamounts are an ample (Menard and Fisher, 1958, Fig. 3). hese in turn grade into closely spaced circular seamounts which are not connected topographically, such as the group associated with Fieberling Guyot (Menard, 1955b, Pl. 1; H.O. Chart 0527).

Small asymmetrical ridges normally are associated with deep troughs, but the ridges around Hawaii are not, nor is one south of the Murray fracture zone. Narrow troughs with parallel asymmetrical ridges on one or both sides have been described both north and south of the Clipperton fracture zone (Menard and Fisher, 1958, p. 249), and others occur in the vicinity of the Clarion, Murray, and Mendocino fracture zones and also south of the Pioneer Ridge which lies parallel to and south of the Mendocino fracture zone.

Straight troughs, attributed to faulting, have been described on the continental sides of the Kurile Trench (Udincev, 1955, Pl. 3) and the Aleutian Trench (Gates and Gibson, 1956, Fig. 3). These troughs superficially resemble normal submarine canyons, but they are straight, have parallel trends, and in both areas they have about the same oblique trend to the trench axis, so that they appear to be fault troughs.

Two patterns of minor lineations are identified, a pinnate pattern centered on major lineations, and a less definite herringbone pattern extending across crustal blocks and changing direction at the fracture zones in the northeastern Pacific. In addition, both major and minor lineations have a relatively constant trend in very large areas near the continental margins of North American and South America.

The pinnate pattern consists of minor lineations on both sides of major lineations with different but relatively constant trends on each side (Fig. 1). The whole pattern resembles a feather with the major lineations as the shaft and the minor lineations as barbs. The angle between representative minor lineations, measured as the "V" of the barbs, ranges from 66° to 140° and commonly is not bisected by the major lineation. Separated into classes according to associated types of major lineation, the angles are either about 70° or 90° for fracture zones, 105°–110° for linear archipelagoes, and about 140° for the Kurile Trench. Although the grouping in classes is suggestive, the sample is

not large enough to assure that significant differences in angles exist in the several types of major lineations.

however, the minor lineations are as uniform in trend as any group of continental faults distributed over a similar area. The lineations

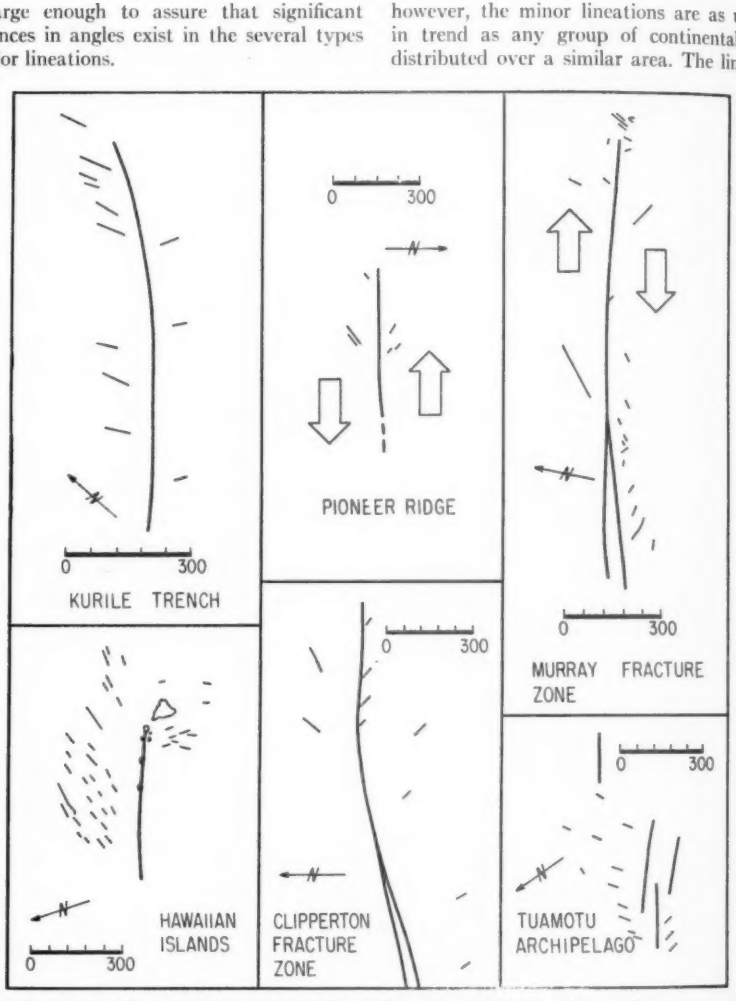

FIGURE 1.—PINNATE PATTERN OF MINOR LINEATIONS AROUND MAJOR LINEATIONS
Scale in nautical miles

An exception to the general rule of constant trend of minor lineations on one side of a major lineation exists south of the Murray fracture zone (Fig. 1) in the Moonless Mountain region (Menard, 1955a, p. 37). These mountains consist of 15 known volcanoes ranging in height to 3500 m, and almost all are markedly linear. (The more nearly circular peaks are not plotted on Figure 1.) The lineations of the peaks which have been surveyed in most detail lie in two groups, one trending northwesterly and the other northeasterly. In most other regions, northeast of the Hawaiian Islands and th south of Clipperton fracture zone (Fig. 1), example, have as constant a trend and spread over about the same area as the fa of the Basin and Range Province in the Uni States.

The herringbone pattern of minor fractu incorporates the pinnate patterns around dif ent fracture zones into a single whole. T arises from the fact that the "V's" in pinnate patterns of adjacent fracture zo point in opposite directions (Fig. 2). S

ost of the minor lineations in the northeastern cific fit into both the herringbone and pin- te patterns the two names, in a sense, merely dicate emphasis, the former on minor linea- lines by various Scripps expeditions indicate lines of closely spaced small seamounts. A similar range of small seamounts between the Murray fracture zone and the Pioneer Ridge

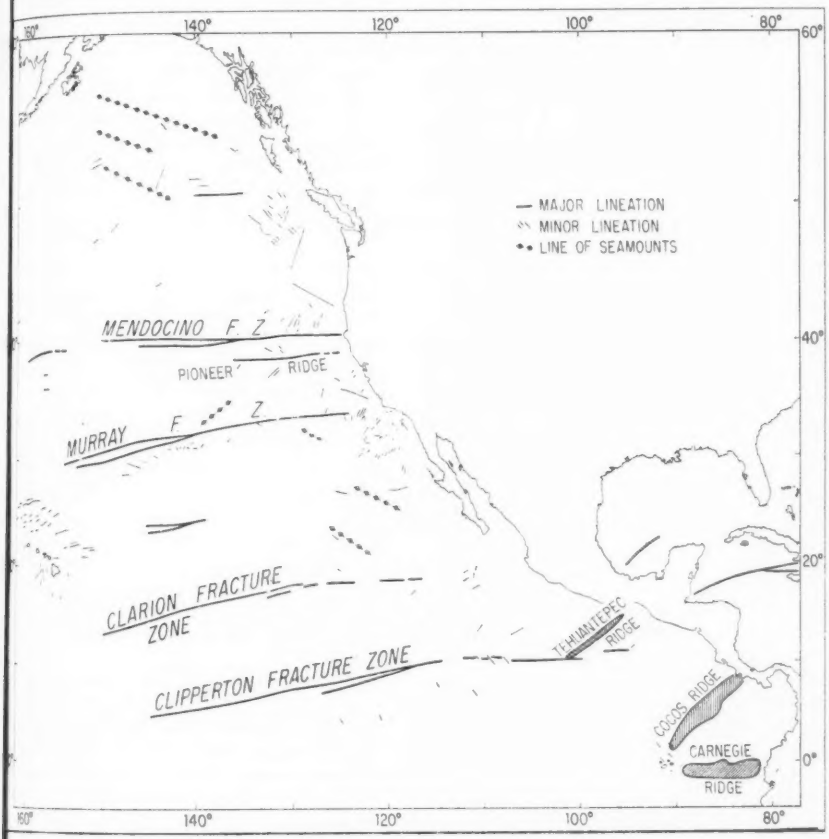

FIGURE 2.—MAJOR AND MINOR LINEATIONS IN THE NORTHEASTERN PACIFIC BASIN
The relatively constant trend of minor lineations between fracture zones forms a herringbone pattern this part of the basin. Off Central America, Mexico, and California, most lineations are northeasterly.

ns within crustal blocks and the latter on nor lineations at the margins of blocks. owever, there are some lineations in the iddle of crustal blocks, suggesting that the rringbone pattern is more significant al- ough it is not so definite as the pinnate one. nong the mid-block lineations, the best ablished are in the Baja California Seamount ovince between the Murray and Clarion cture zones where at least two ranges of sea- unts have the northwesterly trend found at margins. The ranges have not been sur- yed, but closely spaced random sounding trends northeasterly across almost half the width of the crustal block and has the same trend as the fractures at the margins of the block.

A prominent northeasterly trend exists on the deep-sea floor in a band extending from northern California to northern Chile. In the less densely sounded southern region the trend comprises three great submarine ridges, the Tehuantepec Ridge (Menard and Fisher, 1958, Fig. 1), the Cocos Ridge (Shumway, 1954, Fig. 2), and the Nasca Ridge which was surveyed during the DOWNWIND Expedition. All

the ridges are connected with, and appear to branch from, fracture zones with easterly trends. Off Baja California and California, the northeasterly trend is exhibited by at least

tion, 1959) and which may be another of fracture zones of the northeastern Pacific.

Minor lineations consisting of elongate canoes, lines of volcanoes, asymmetrical ri

FIGURE 3.—COMPARISON OF THE SCALE OF MAJOR AND MINOR LINEATIONS AROUND THE HAWAIIAN ISLA (mirror image) WITH THE SAN ANDREAS FAULT AND THE BASIN RANGES

20 large volcanoes including some on the continental slope and borderland (Shepard and Emery, 1941, p. 33–34). The northeasterly lineations have not been found north of Pt. Arena in California where the San Andreas fault trends out to sea. A much less prominent northerly trend of minor lineations also exists off Baja California. This trend includes the active volcano of San Benedicto Island, the undissected volcanoes of Guadalupe Island, and a few small seamounts. Inasmuch as none of the volcanoes with northeasterly trends is active, these with northerly trends may be younger.

A well-developed group of lineations off Vancouver trends northwesterly. Individual lineations are parallel to and lie in the area between the active seismic lines northwest of Cape Mendocino and southwest of Queen Charlotte Island (Gutenberg and Richter, 1954, p. 34). The group also lies southeast of an east–west ridge which has been traced for 350 km (Robert Hurley, personal communica-

and narrow troughs mark tension faults. T origin of the tension, however, need not uniform. The en echelon pattern of a group very elongate volcanoes off Baja Californ suggests tension cracks produced by an unde lying shear. The minor lineations around t Hawaiian Islands in general lie on the Hawai Arch (Dietz and Menard, 1953, Fig. 3; Ham ton, 1957, Fig. 4), and they may be formed keystone faulting incidental to the uplift the arch. Likewise, most troughs have asy metrical ridges on each side and conceiva might result from keystone faulting on a clines. However, volcanoes, asymmetr ridges, and troughs all have the same orien tion in some regions, and it is difficult to beli that they do not have a common origin.

The oceanic crust of the eastern Pacific fractured into blocks with dimensions of t order of 10^5–10^6 km^2. Major easterly lineati called fracture zones (Menard, 1955b) exist the margins of the blocks, and large displa ments have taken place on some of, if not

fracture zones. Right-lateral faulting of bout 130 km has been measured on the Murray acture zone (Mason, 1958, p. 328; Menard nd Vacquier, 1958, p. 2–5), and left-lateral ulting of 250 km on the Pioneer Ridge (Victor acquier, personal communication, 1959). ominate minor lineations border the two frac- re zones where displacement has been meas- ed (Fig. 1). A similar pattern of minor linea- ons is associated with many other major ineations in the Pacific Basin. If the major and inor lineations could be related to a common rigin, it would suggest that horizontal dis- acement has occurred on all the major linea- ons.

The herringbone pattern is not definite nough at present to justify elaborate specula- ons about origin. The strain pattern seems to e constant within a crustal block and differs stematically from block to block, but even ese conclusions need confirmation from rther mapping.

The scale of the crustal distortion implied y the common association of patterns of minor ineations with major lineations merits atten- on. If a mirror image of the Hawaiian Islands ineation is compared with the San Andreas ult (Fig. 3), the minor lineations associated ith the Hawaiian lineation lie in the same area s the Basin Ranges and have about the same ends, spacing, and lengths as individual anges. All these similarities may be attributed o coincidence, but the scale is significant. orizontal displacement along the San Andreas ult of the magnitude suggested by Hill and ibblee (1953, p. 445–449) implies major rustal shortening and stretching somewhere. he distribution of major and minor lineations n the sea floor suggests that geologists may stifiably look far afield for the secondary ffects of large-scale wrench faulting.

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MANUSCRIPT RECEIVED BY THE SECRETARY OF THE SOCIETY, MAY 15, 1959

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